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CONTENTS

	PAGES
Kotlassia prima Amalitsky. By A. P. Bystrow	379–416
Gravity investigation of central-eastern Pennsylvania. By J. B. Hersey	417–444
Igneous rocks of the northern Big Belt Range, Montana. By John B. Lyons	445–472
Magmatic contrasts in the Winnipesaukee region, New Hampshire. By Alonzo Quinn	473–496
Petrology and structure of the Oliverian magma series in Mt. Washington quadrangle, New Hampshire. By Carleton A. Chapman, Marland P. Billings, and Randolph W. Chapman	497–516

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Correlation Charts Prepared by the Committee on Stratigraphy of the National Research Council, Carl O. Dunbar, *Chairman*

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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KOTLASSIA PRIMA AMALITZKY

BY A. P. BYSTROW

CONTENTS

	Page
Abstract	380
Introduction	380
Description	381
Cranial structure	381
Skull topography	381
Skull roof	381
Palatal surface	382
Braincase and stapes	384
Lower jaw	389
Tooth structure	390
Vertebral column and ribs	392
Appendicular skeleton	399
Restoration	406
Armor	407
Discussion	407
Synonymy of described species	407
Relationships	411
Environment	412
Comparison with *Seymouria*	413
Postscript	415
References cited	415

ILLUSTRATIONS

Figure	Page
1. Dorsal view of the skull of *Kotlassia prima*	382
2. Ventral view of the skull of *Kotlassia prima*	383
3. Pterygoid and neurocranium of *Kotlassia prima*	385
4. Occipital view of skull; neurocranium	386
5. Anterior view of the neurocranium; sphenethmoid; and stapes	388
6. Lower jaw of *Kotlassia prima*	389
7. Sections of teeth of *Kotlassia prima*	391
8. Vertebrae and ribs of *Kotlassia prima*	392
9. Vertebrae and ribs of *Kotlassia prima*	393
10. Vertebrae and ribs of *Kotlassia prima*	394
11. Vertebrae and ribs of *Kotlassia prima*	395
12. Vertebrae and ribs of *Kotlassia prima*	396
13. Vertebrae and ribs of *Kotlassia prima*	397
14. Vertebrae and ribs of *Kotlassia prima*	398
15. Pectoral girdle of *Kotlassia prima*, ventral view	400
16. Pectoral girdle and limb of *Kotlassia prima*	401
17. Pelvis of *Kotlassia prima*, lateral view	402
18. Pelvis of *Kotlassia prima*, ventral and dorsal views	403

19. Pelvic limb of *Kotlassia prima* 404
20. Restoration of the skeleton of *Kotlassia prima* 405
21. Armor of *Kotlassia prima* 408
22. Vertebrae and ribs of *Kotlassia prima* (*Kotlassia secunda* Amalitzky = *Karpinskiosaurus secundus* Sushkin) 410

ABSTRACT

The skeletal material of *Kotlassia prima*, a primitive tetrapod from the Upper Permian of the North Dvina River in northern Russia, is described and figured in detail. The various specimens permit a nearly complete description except for the feet. *Karpinskiosaurus secundus* is a synonym of *Kotlassia prima*. The animal is definitely amphibian in nature although it may be regarded as a relict of the reptilian ancestors of the Reptilia. *Kotlassia* was as an adult a terrestrial form, in contrast to the contemporary *Dvinosaurus*. *Seymouria* may be an amphibian of the same kind.

INTRODUCTION

At the beginning of the present century (1899–1910), V. P. Amalitzky discovered in the Upper Permian deposits of the basin of the North Dvina River two incomplete skeletons, two skulls, and isolated disconnected bones of "Seymouridae". They were situated in a lens of sand with skeletons of Pareiasauria, Inostrancevia, Dicynodontia, Dvinosauria, and other representatives of the so-called North Dvina fauna.

Amalitzky (1921) first described these materials and found a great analogy between the forms of North American Seymouridae and those found in the North Dvina deposits. He concluded that they belonged not only in the same family (Seymouridae), but even in the genus *Seymouria* in the order Cotylosauria. However, taking into consideration certain morphological peculiarities—the differences in age of the deposits, and the great distance between the distribution areas—Amalitzky considered it necessary to separate the Upper Permian, North Dvina *Seymouria* as an isolated subgenus *Kotlassia*. Two incomplete skeletons differed somewhat from one another so that he described them as two species of this subgenus: *Kotlassia prima* and *Kotlassia secunda*.

Later Sushkin (1925; 1927; 1928), after a slight additional preparation of the skull of *Kotlassia prima*, No. 74,[1] again described the material of Amalitzky. On the grounds of peculiarities in the structure of stapes, quadratum and pterygoideum, he concluded that both *Kotlassia* and *Seymouria* of North America, are Amphibia and not Reptilia. They were considered reptilelike by many palaeontologists including Broili (1904; 1916; 1927), Case (1911; 1915), Huene (1912; 1913), Watson (1914; 1916; 1918), and Williston (1910; 1911; 1914; 1925). Sushkin changed the name of *Kotlassia secunda* to *Karpinskiosaurus secundus*, considering it to be the representative of a new genus and a new family (Karpinskiosauridae).

In 1929, A. P. Hartmann-Weinberg discovered in the collections of Amalitzky a fragment of armor which, according to her, belonged to *Kotlassia*. By studying the microscopic structure of this armor, Hartmann-Weinberg endeavored to clarify the systematic position of *Seymouria* in general. However, in her paper of 1935, she was unable to decide whether the microstructure of the armor in the Russian Seymouriamorpha was amphibian or reptilian. Hartmann-Weinberg did not study the skeletons of Seymouridae.

[1] These numbers refer to specimens in the collections of the Paleozoological Institute of the Academy of Sciences of the U.S.S.R.

The three investigations mentioned (Amalitzky, Sushkin, Hartmann-Weinberg) constitute all the literature we possess on the *Kotlassia* of the North Dvina fauna. They cannot, in any case, pretend to give a complete description of the morphological peculiarities of these interesting forms. An examination of the authentic skeletons in the collection of Amalitzky convinced the author that these skeletons required a very thorough additional preparation before they would be suitable for a sufficiently detailed study.

Such a preparation has been made of the skeleton of *Kotlassia prima* (No. 74) and in somewhat lesser measure of the skeleton of *Kotlassia secunda* (*Karpinskiosaurus secundus*) (No. 81). In skeleton No. 74, the skull, the ventral surface of the vertebral column, a considerable part of the ribs, and the pelvis have been prepared; in skeleton No. 81, the lower jaw, quadratum, and certain vertebrae. Formerly all this was so covered by matrix that it was impossible to give a detailed description. The removal of the matrix made it possible to study previously unknown details of structure in the skeleton.

The results of this study are given in the present paper. All the illustrations have been prepared by the author as accurately as possible. The strong deformation of the *Kotlassia* skulls has been rectified on the drawings.

DESCRIPTION

CRANIAL STRUCTURE

Skull topography.—By its general configuration the skull of the North Dvina *Kotlassia* must be placed among the short and wide types. The length of the skull along the median line varies in the three specimens (Nos. 74, 81, 82) from 7.5 to 12.5 cm.; the width from 8.5 to 14 cm. The external nares in the skull have been moved to its anterior edge; the orbits are situated rather far apart; a moderately large pineal opening exists in all the skulls; the otic notch is very deep; the horns of the tabular are directed backwards and sideways.

Skull roof (Fig. 1).—The dermal bones which form the roof of the skull of *Kotlassia* are placed as in the typical labyrinthodonts. All these bones are covered by a round-celled sculpture typical of the labyrinthodonts. Only in the parts in which there is the most intense growth do the round cells on some bones take the shape of long furrows (parietal, squamosal, jugal, quadratojugal). The furrows of the lateral-line system are entirely absent.

PREMAXILLARY: A short and broad bone, situated on the blunted anterior edge of the skull. By its posterolateral edge it bounds the oval external naris. It has been possible to uncover a small interpremaxillary foramen between the premaxillaries.

NASAL: Small, but considerably larger than premaxillary.

FRONTAL: About twice as large as the nasal. The dimensions of the parietal are still greater. In this way, the size of the dermal bones on both sides of the median suture of the skull rapidly increases from the front backwards. Behind the parietal, however, comes a pair of comparatively small dermo-supraoccipitals, of unequal dimensions.

Five bones form the limits of the orbits of *Kotlassia*: lacrimal, pre- and postfrontal,

postorbital, and jugal. The first four bones are comparatively small; the jugal is somewhat larger. The outlines and situation of the lacrimal are such that this bone forms a part of the border of the orbit and of the external naris.

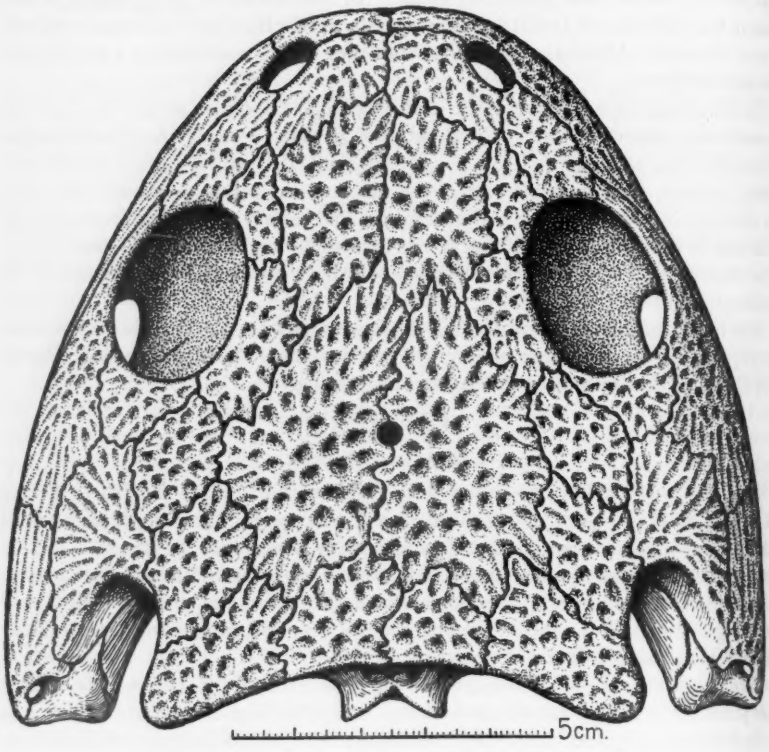

FIGURE 1.—*Dorsal view of the skull of* Kotlassia prima
Natural size.

Behind the postfrontal and postorbital a small intertemporal is situated in the roof of the skull. The deep otic notch penetrates so far into the skull roof that there is no contact between squamosal and tabular, and the border of the otic notch is formed for a considerable length by the edge of the supratemporal.

Palatal surface (Fig. 2).—Along the palatal edges of premaxillary and maxillary is situated a row of conical, slightly curved teeth. In transverse sections, their bases are oval-shaped, the long axes of the ovals being placed perpendicular to the external edge of the skull.

Behind each premaxillary is a rather large vomer, which, on the posterolateral side, limits a small oval-shaped internal naris. Behind the internal naris, adjoining on the inside the long and narrow maxillary, are two bones, one after another: palatine

and transverse. Teeth could be discerned on both bones. They form a slightly S-shaped curved row. The size of these teeth, beginning at the internal naris, decreases gradually. Among them there are no so-called "tusks", which occur on these bones in labyrinthodonts.

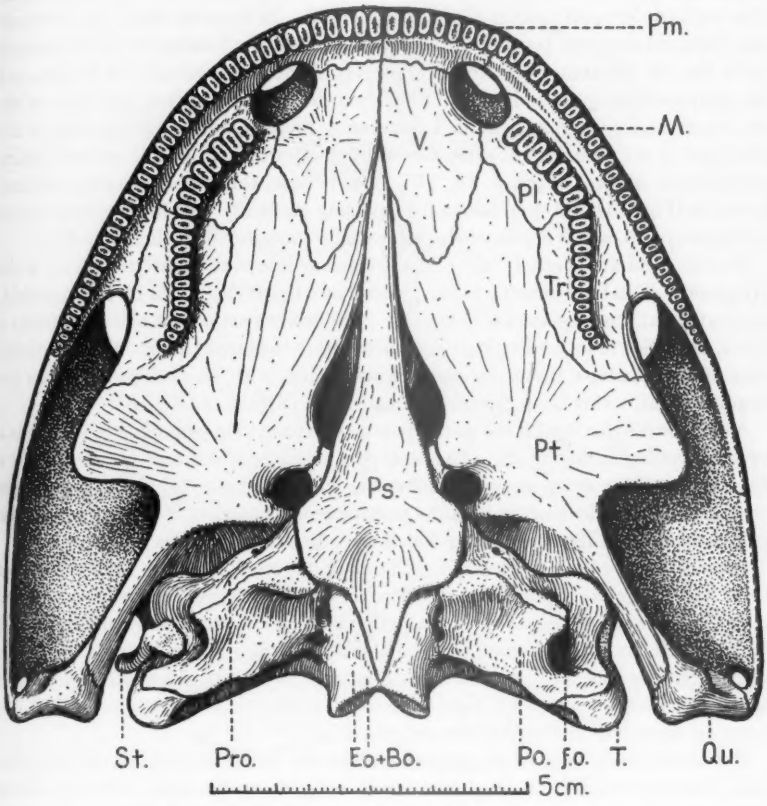

FIGURE 2.—*Ventral view of the skull of* Kotlassia prima

Bo—basioccipital; Eo—exoccipital; f.o.—fenestra ovalis; M—maxillary; Pl.—palatine; Pm.—Premaxillary; Po—paroccipital; Pro.—pro-otic; Ps—parasphenoid, Pt—pterygoid; Qu—quadrate; T—tabular, Tr—transverse; St—stapes; V—vomer. Natural size.

The pterygoids of *Kotlassia* have a complicated shape. The anterior branch of every pterygoid is very broad. Being situated in a horizontal plane, it constitutes a considerable part of the bony palate. The anterior end of the pterygoid, forming two large projections, penetrates between the processus cultriformis of the parasphenoid and the vomer, and between the vomer and palatine. Besides this the pterygoid has a broad blunted lateral outgrowth, which joins the transverse behind, narrowing considerably the anterior part of the infratemporal vacuity. The two

pterygoids are drawn so close together by their anterior branches that between them and the processus cultriformis there remain but very narrow interpterygoid spaces. To the front these spaces pass into narrow slits, which soon disappear entirely, owing to a complete contact between the pterygoid and the parasphenoid. Posteriorly, the interpterygoid spaces are closed by the union of a special process on the pterygoid with the basipterygoid process of the basisphenoid. In forms in which the pterygoid was displaced after the death of the animal, but before the fossilization of its skeleton (as in No. 74) the contact between the pterygoid and basisphenoid was broken, and the pterygoid easily pushed aside. This, doubtlessly, is proof that the union of the two bones was not strong. Behind this union, there is on the median edge of the pterygoid a semicircular incisure, directly behind which a well-developed lamina ascendens is situated (Figs. 2, 3A, 3B). At its base, this plate is slightly inclined forwards (Fig. 3A) and then backwards, passing gradually into the posterior ramus of the pterygoid, which is groove-shaped, with its concave side facing laterally.

A small quadrate (Figs. 2, 3C) joins the concave side of the posterior ramus of the pterygoid. This bone penetrates like a wedge into the angle formed by the pterygoid, quadratojugal, and squamosal (Fig. 4A). The lower part of the quadrate forms a block, expanded in its median part and narrowed in the lateral one. On the external edge of the quadrate there is a small notch which, with an analogous one on the quadratojugal, forms the foramen paraquadratum (Figs. 1, 2, 3C, 4A).

Along the median line on the skull's palatal surface a long parasphenoid presents a very interesting form (Fig. 2). The basal part of this bone is drawn backward, like a flat wedge, whose sharp summit reaches the edges of the occipital condyles. The ventral surface of the basal part of the parasphenoid is concave. The long processus cultriformis lies at its extremity on the dorsal edges of both vomers.

Braincase and stapes (Figs. 4, 5).—The occipital condyle of the skull consists of three bones: two exoccipitals and a basioccipital. These three bony elements, forming an occipital condyle, have grown together to such a degree that no distinct limits between them can be discerned. The articulated surface of the occipital condyle is rather strongly concave. A small ossified basioccipital forms the middle part of this surface. It has been possible to prepare in the center of the basioccipital (Fig. 4A) a certain stretch of the canal of the notochord.

The exoccipitals, their bases concrescent with the lateral surfaces of the basioccipitals, pass upward and, limiting the foramen magnum on the sides, unite by sutures with the dermosupraoccipitals. The lower edge of the foramen magnum is formed by two flat processes of the exoccipitals (processus submedullaris). The sharp tips of these processes meet one another above the basioccipitals in such a fashion that an opening remains between them (fenestra basioccipitalis), which apparently was filled by cartilage.

Lateral to the exoccipitals are powerful paraoccipitals. Between the dorsal process of the exoccipital and paraoccipital, is a large foramen, which served for the passage of the IXth and Xth cranio-cerebral nerves (Fig. 4A). Below this foramen the bones (exoccipital and paroccipital) are united by a simple suture, in which there are, however, rather wide fissures. These fissures are not symmetrical on the two halves of the skull. Apparently they were filled with cartilage and did not serve for the exit

of any cranio-cerebral nerves, or blood vessels (Figs. 2, 4A). There are no other openings in the exoccipitals placed more caudally than the exit of the IXth and Xth nerves. This proves that, in *Kotlassia*, N. hypoglossus (XII) did not pass out of the

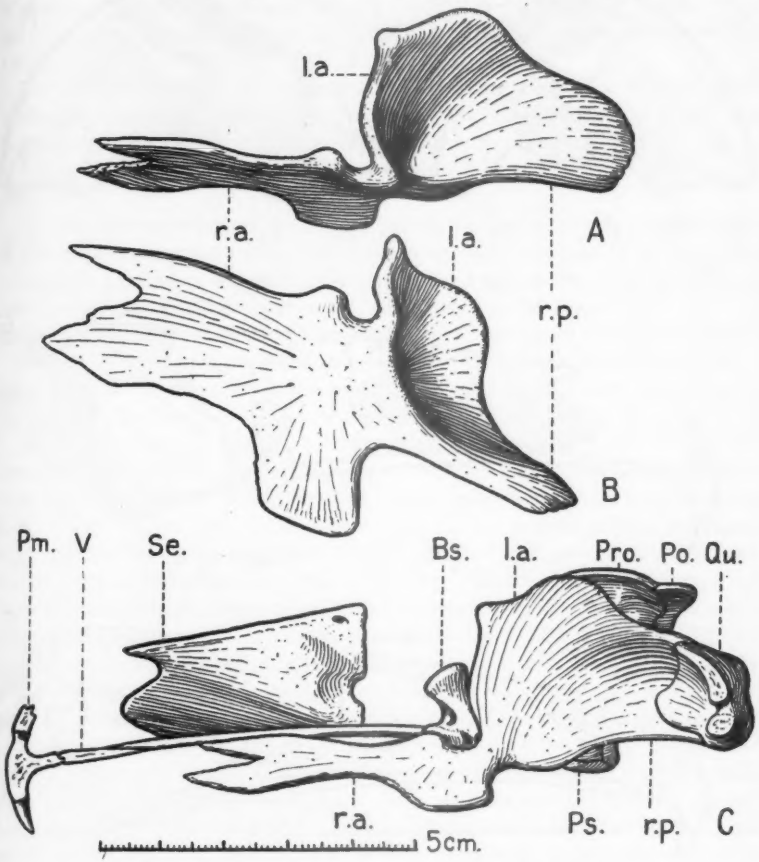

FIGURE 3.—*Pterygoid and neurocranium of* Kotlassia prima

A—medial surface; B—ventral surface of the pterygoid; C—Neurocranium and pterygoid. Bs—basisphenoid; l.a.—lamina ascendens; Pm—Premaxillary; Po—Paroccipital; Pro—pro-otic; Ps—parasphenoid; Qu—quadrate; r.a.—anterior branch; r.p.—posterior branch of the pterygoid; Se—sphenethmoid; V—vomer. Natural size.

skull by a special foramen but left the cerebral canal of the axial skeleton at the back of the exoccipitals.

The paroccipitals by their inferior blunted angles descend ventrally so far that they reach the parasphenoid (Figs. 2, 4A). The upper-medial angle of the paroccipital joins the dermosupraoccipital by a suture, forming the anterolateral edge of the foramen of the IXth and Xth nerves. The upper lateral angle of the paroccipital

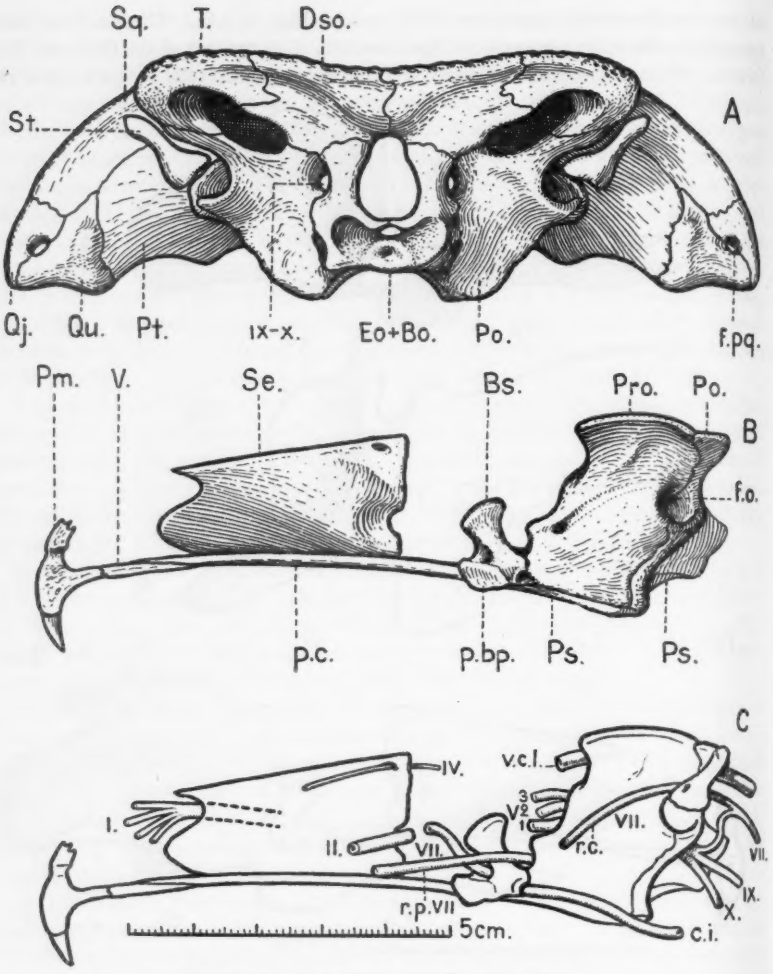

FIGURE 4.—*Occipital view of skull; neurocranium*

A—Occipital view of the skull of *Kotlassia prima*; B, C—Neurocranium of *Kotlassia*. Bo—basioccipital; Bs—basisphenoid; Dso—dermo-supraoccipital; Eo—exoccipital; f.o.—fenestra ovalis; f.p.q.—paraquadrate foramen; p.c.—processus cultriformis; p. bp.—basipterygoid process; Pm—premaxillary; Pro—pro-otic; Po—paroccipital; Ps—parasphenoid; Pt—pterygoid; Qj—quadratojugal; Qu—quadrate; Se—sphenethmoid; Sq—squamosal; St—stapes; T—tabular; V—vomer. Natural size.

is drawn out as a long flattened process, upon the end of which the tabular is placed. Thus in the occipital part of the skull an elongated opening (fenestra subtabularis) is formed, limited by tabular, dermosupraoccipital, and paroccipital. Directly under the process by which the paroccipital is joined to the tabular, the lateral edge of the paroccipital bounds the wide fenestra ovalis posteriorly.

On one of the skulls of the North Dvina *Kotlassia* (No. 74) the stapes has remained on the right side. This bone has been described in its general features by Sushkin (1927). Sushkin did not see in it a foramen for the stapedial artery and considered it absent. However, additional preparation makes this opening visible. It has not been completely preserved, as one of the edges of the stapes was destroyed during the first preparation (Fig. 5E).

The stapes of *Kotlassia* is a comparatively small ossicle with a widened massive base and a somewhat thickened summit. On the exterior edge of the bone a longitudinal, slightly bent, blunt crest thickens at the summit and at the base. In front of this crest in the immediate vicinity of the broad base, the stapes is pierced by a small canal.

The widened lower part of the stapes does not adhere to the edges of the fenestra ovalis indicating that the interval between the stapes and the edge of the paroccipital behind was covered at the back by a cartilaginous operculum, as in contemporary Anura and Urodela. The upper end of the stapes, directed toward the center of the otic notch, was, without doubt, united here to a well-developed tympanic membrane.

In front a large bony pro-otic unites with the paroccipital (Figs. 2, 4B, 5A). It participates in the formation of the fenestra ovalis bounding it anteriorly (Fig. 4B). By its lower edge the pro-otic is joined to the parasphenoid, while by the upper surface it reaches the internal side of the covering bones of the skull roof. Here, inward from the otic notch, the pro-otic becomes considerably wider and is surrounded in front by the edge of the ascending plate (lamina ascendens) of the pterygoid (Figs. 2, 3C, 5A). On the lateral surface of the pro-otic some distance from its anterior edge, one can see an opening, facing backwards and upwards, from which goes a shallow, but quite noticeable, furrow (Fig. 4B). Without doubt, the ramus communicans nervi facialis, which came out of the indicated opening, lay here and passing over the stapes united with Nervus glossopharyngeus (IX) (Fig. 4C). Upwards and backwards from the median side of the pro-otic extends a narrow and long process (Fig. 5A). At its summit it reached the small but sufficiently well-ossified supraoccipital. This bone can only be seen from the side of the cerebral cavity of the skull, as it is completely covered behind by the dermosupraoccipital. Between the much-widened lateral and the long narrow medial process of the pro-otic there exists, on the upper surface of this bone, a small notch which, with the covering bones of the skull roof forms a slit. Through this slit the vena capitis lateralis (v.c.l., Fig. 4C) passed out of the fenestra subtabularis. In the lower section of the anterior edge of the pro-otic there is a small notch. Here the ramus palatinus nervi facialis (r-p, VII, Fig. 4C) came out on the palate. Somewhat higher, and more medially than the anterior edge of the pro-otic, three branches of the trigeminus nerve leave the cerebral cavity (V_1, V_2, V_3, Fig. 4C).

On the parasphenoid, directly adjoining the anterior edges of both pro-otics, is the bony basisphenoid. This bone has a well-developed crest—the dorsum sellae turcicae, somewhat inclined forward. On the front surface of this crest is a round hole, in which the hypophysis was situated (Fossa hypophyseos—f.h., Fig. 5A). The lateral parts of the basisphenoid extend outward beyond the edges of the parasphenoid and form the basisphenoid processes to which the pterygoids are joined. At the

base of these basisphenoid processes there are short canals, through which passed the internal carotid artery (c.i., Fig. 4C).

A little to the front of the basisphenoid, contiguous at its lower edge with processus

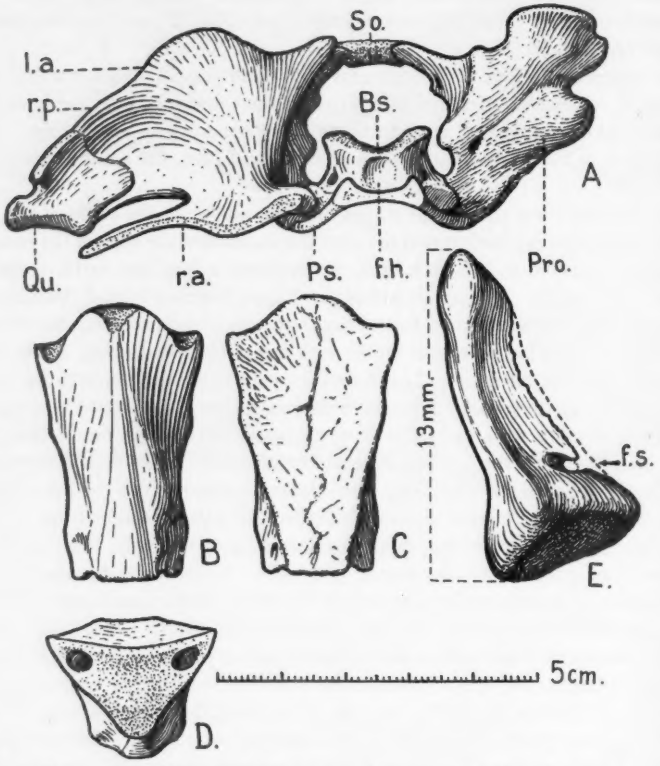

FIGURE 5.—*Anterior view of the neurocranium; sphenethmoid; and stapes*

A—Anterior view of the neurocranium of *Kotlassia*; B—sphenethmoid, inferior view, C—superior view, D—anterior view of the sphenethmoid; E—lateral view of the stapes. Bs—basisphenoid; f.h.—fossa hypophyseos; f.s.—foramen of stapedial artery; l.a.—lamina ascendens; Pro—pro-otic; Ps—parasphenoid; Qu—quadrate; r.a.—anterior branch of the pterygoid; r.p.—posterior branch of the pterygoid; So—supraoccipital. Natural size.

cultriformis, is a well-ossified sphenethmoid (Figs. 3C, 4, 5). This bone is shaped like a trihedral prism and differs little from the sphenethmoids of *Wetlugasaurus* and *Zygosaurus*. The upper, slightly convex surface of the sphenethmoid is widened in the anterior section and narrowed in the posterior (Fig. 5C). The lateral surfaces of the bones meet at an angle with the processus cultriformis (Fig. 5B, D). In the lower section of the posterior edge of the sphenethmoid there is a small notch, which limits in front the foramen opticum (II, Fig. 4B, C). Above it at the very upper margin of the sphenethmoid is situated a very small oval-shaped opening, which served for the exit of the very fine Nervus trochlearis (IV, Fig. 4B, C). The author

discovered a long stretch of two canals through which the olfactory nerves (I, Fig. 4B, C) passed in the substance of the sphenethmoid.

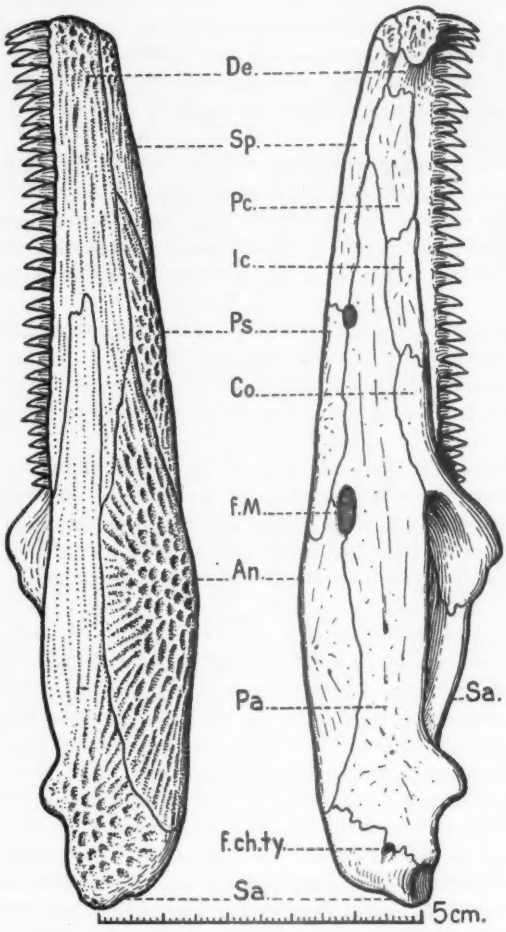

FIGURE 6.—*Lower jaw of* Kotlassia prima

An—angular; Co—coronoid; De—dentary; f.M.—foramen Meckelianum posterius; f.ch.ty—foramen chordae tympani; Ic—intercoronoid; Pa—prearticular; Pc—precoronoid; Ps—postsplenial; Sa—supra-angular; Sp—splenial. Natural size.

Lower jaw (Fig. 6).—The material at the disposal of Amalitzky and Sushkin was so badly prepared that the lower jaw of *Kotlassia* could not be sufficiently studied. The further preparation of objects Nos. 74 and 81 enables the author to give a complete description of the structural details of the lower jaw.

A small bony articular lying in the most posterior section of the lower jaw is almost completely encased externally, posteriorly, and internally by a well-developed supra-angular (Fig. 6). The external surface of the supra-angular in its posterior part is covered by the usual round-celled sculpture; in front, the bone is drawn out in the shape of a long wedge, more than half the length of the lower jaw.

On the internal side a still longer covering bone—the prearticular, adjoins the articular. The supra-angular and prearticular cover the articular so that it can be seen only from above.

In the suture dividing supra-angular and prearticular, there is a small aperture, which served as a point of entrance of chorda tympani (f.ch.ty., Fig. 6) into the lower jaw. The lower angle of the jaw is formed by a large angular, joined by sutures to supra-angular and prearticular. All the external surface of the angular, as far as the lower edge of the jaw, is covered by a cellular sculpture; the internal surface of the bone is smooth.

In the anterior part of the suture between angular and prearticular there is a rather large oval-shaped foramen—Meckelianum posterius.

The long dentary is adjoined posteriorly and overlapped by supra-angular and angular. There is a row of conical slightly curved sharp teeth all along the upper edge of the dentary.

Along the lower edge of the jaw from its symphysial end to the angular lie two bones, one behind the other—splenial and postsplenial. Both are covered by a cellular sculpture on the outside and are quite smooth on the inside.

On the inside of the lower jaw, at the point where the sutures separating splenial, postsplenial, and prearticular meet, there exists a small foramen—Meckelianum anterius.

Between the dentary and the long anterior outgrowth of the prearticular, three toothless coronoids can be seen in the lower jaw of *Kotlassia*. The coronoid is a great deal more developed than intercoronoid and precoronoid. Its posterior part forms a high crest, closely adhering to the inside of the supra-angular. The edge of this crest is strongly thickened, and when the jaws are shut on the whole skull it enters into that narrow part of the fossa infratemporalis which is limited on the medial side by the protuberance of the anterior branch of the pterygoid (Fig. 2).

The general structure of the lower jaw of *Kotlassia* does not differ essentially from the structure of the jaw of typical labyrinthodonts. The presence or absence of teeth on the coronoids is exceedingly variable among the labyrinthodonts. Thus in *Eogyrinus*, *Orthosaurus*, *Eryops*, *Buettneria* (Broom, 1913; Case, 1932; Watson, 1926; Williston, 1914) teeth are lacking on all the coronoids as in *Kotlassia*; in *Benthosuchus* and *Trematosaurus* teeth exist on only one coronoid; in *Trimerorhachis* coronoid and intercoronoid have teeth (Case, 1935); and in *Dvinosaurus* teeth exist on all the coronoids (Sushkin, 1936; Bystrow, 1938).

Tooth structure (Fig. 7).—Fragmentary material in the collection of Amalitzky has enabled the author to prepare two cross sections of teeth of *Kotlassia* for microscopic study. On other objects, which did not allow a preparation of transparent cross sections, the teeth were gradually cut away transversely from top to bottom, and each section has been studied under a magnifying glass. This enabled the author to form a sufficiently distinct picture of their structure.

The dentinal wall of the teeth in *Kotlassia* is folded in the same measure as in the typical labyrinthodonts, for instance, in *Benthosuchus sushkini* (Bystrow, 1938). Teeth of *Benthosuchus* and *Kotlassia* of equal dimensions have the same number of

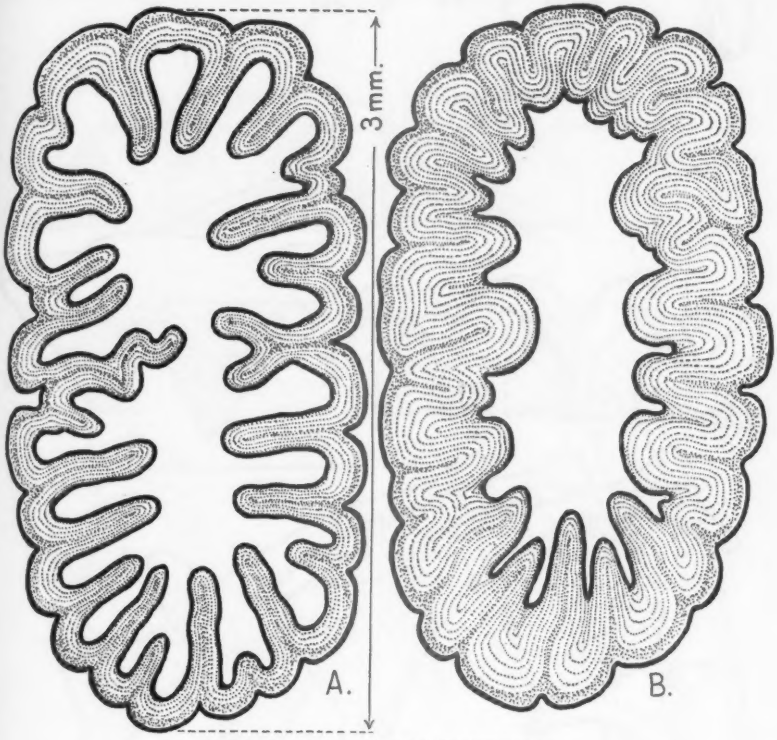

FIGURE 7.—*Sections of teeth of* Kotlassia prima
× 39 approx.

folds. The paucity of material available for preparing the cross section did not allow the author to follow in detail all the stages of the thickening in the dentinal wall, the processes of resorption, and the order of change of the teeth of *Kotlassia*, all of which has been traced in the dental system of *Benthosuchus*. However, one may suppose that in *Kotlassia* all these phenomena took place in the same manner as in a typical labyrinthodont. A study of the slides in which there were several teeth in a row made it possible to discern that at the side of every comparatively thin-walled tooth (Fig. 7A) there is usually a tooth with very thick walls (Fig. 7B).

The investigation of the teeth of *Benthosuchus* has shown that the thin-walled tooth is a young one, the thick-walled, an old one. The distribution of young (Fig. 7A) and old (Fig. 7B) teeth on the jaw of *Kotlassia*, one near another, doubtlessly testifies to the same alternating periodicity in the change of teeth as has been proved for *Benthosuchus*.

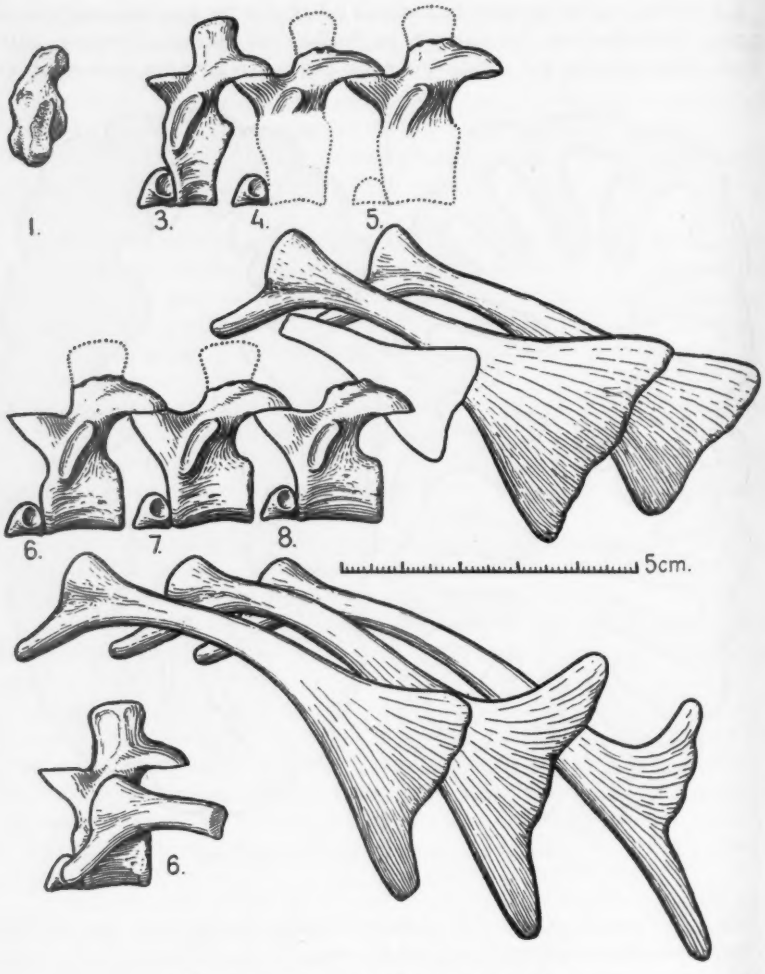

FIGURE 8.—*Vertebrae and ribs of* Kotlassia prima
× ½.

All the teeth of *Kotlassia* belong to the type of the so-called "small" teeth. Large "tusks" are absent on the palatal bones of the skull and on the lower jaw.

VERTEBRAL COLUMN AND RIBS

The vertebral column (Figs. 8–14, 22) in *Kotlassia prima* (No. 74) is almost complete. Some elements of the first vertebra, the entire second vertebra, and 10–15 of the caudal ones are lacking in it. The last-preserved vertebra is the 55th. The rib of the third vertebra remains only as an incomplete impression on the rock; the

remaining ribs are complete on one side of the skeleton; on the other side, the anterior ones have been destroyed.

The additional preparation of skeleton No. 74 has allowed us to study with sufficient accuracy not only the structure of the vertebrae but of all the ribs.

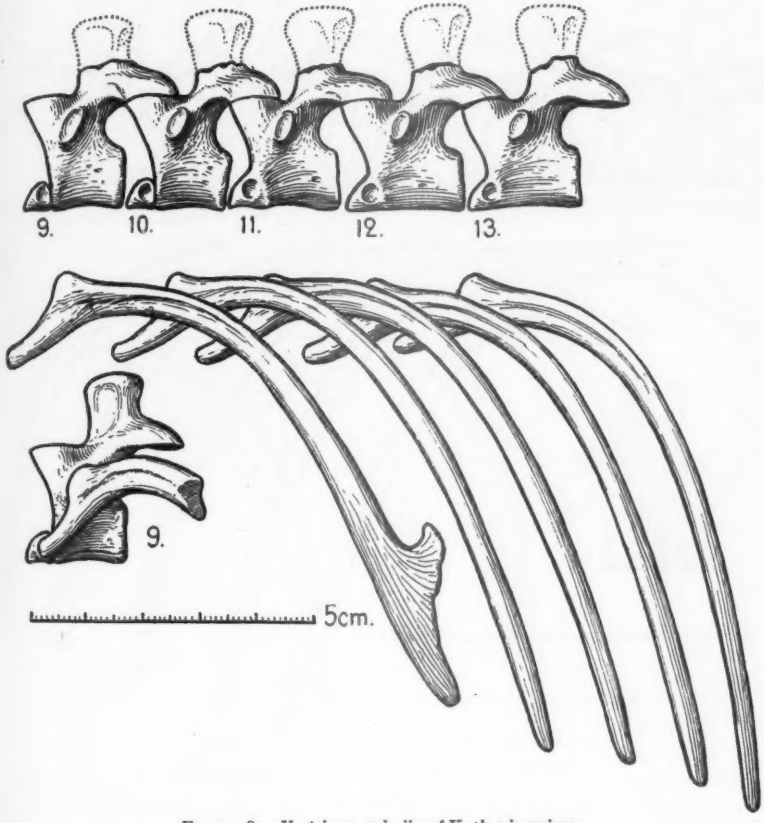

FIGURE 9.—*Vertebrae and ribs of* Kotlassia prima
× $\frac{5}{16}$.

In *Kotlassia prima* (No. 74) only half the neural arch of the atlas (Fig. 8) remains. It represented certainly a quite independent formation and was not concrescent with a like formation on the opposite side. The first three vertebrae are well preserved in *Kotlassia secunda* (*Karpinskiosaurus secundus*). The neural arch of the atlas consists here also of right and left halves. The intercentrum (hypocentrum) of the atlas is represented by two independent bones, this, in all probability, being due to the youth of the given specimen of *Kotlassia*. The second vertebra of *K. secunda* has a high neural spine bent forward. The neural arch of this vertebra is completely fused with the centrum (Fig. 22).

In *K. prima* (No. 74) and *K. secunda* (No. 81) no ribs were joined to the atlas.

The second vertebra (*K. secunda*, No. 81, Fig. 22) also bore no ribs. The third vertebra in both skeletons is very well preserved (Figs. 8, 22). In the additional preparation it has been completely separated from the matrix, and can be described

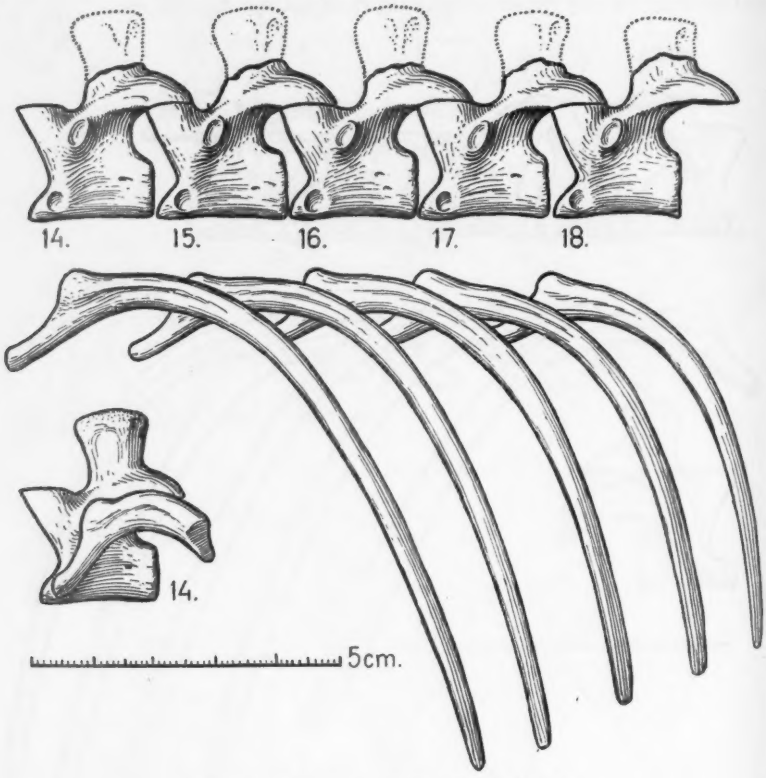

FIGURE 10.—*Vertebrae and ribs of* Kotlassia prima
Natural size.

as an isolated formation in *K. prima*. Its neural arch has completely fused with the centrum; the neural spine is comparatively weakly developed. Where the neural arch unites with the body of the vertebra strong transverse processes go off laterally; the articular surfaces for the tubercle of the rib, on their extremities, have the outlines of slightly curved ovals, narrowed a little ventrally (3a, Fig. 14). The body of the vertebra is doubly concave (amphicoelous) (3b, 3c, Fig. 14); a narrow canal for the notochord can be well seen in its center. The hypocentrum is a small bone, approximately semilunar in shape, situated immediately before the lower edge of the vertebral body (3a, 3b, Fig. 14). The lower surface of the hypocentrum is smooth and more or less flat, the upper surface, concave and rough. On the lateral surface of

the hypocentrum a hollow is seen for the capitulum of the rib (3a, Fig. 14). The third vertebra certainly bore a well-developed rib. Unfortunately, we possess only an impression of its distal half (Fig. 8).

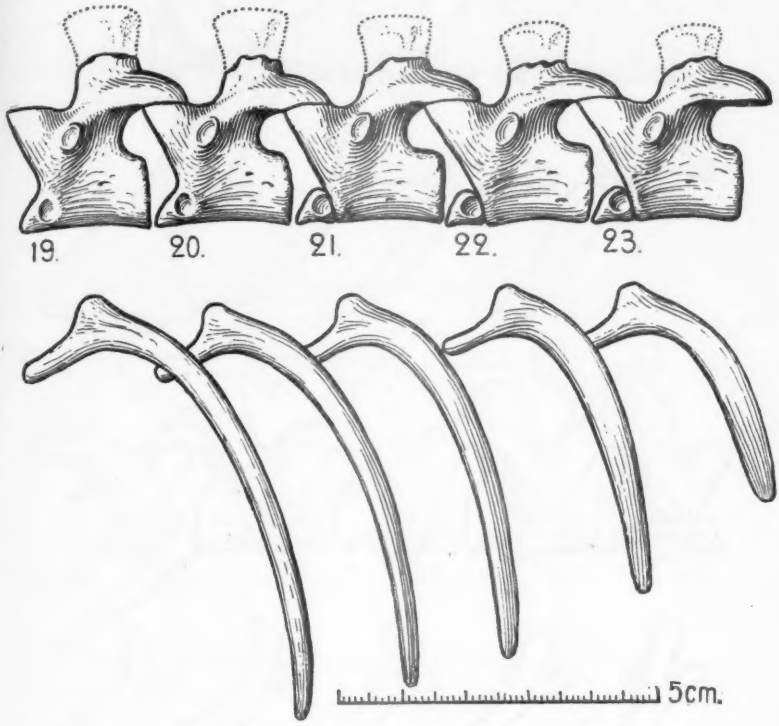

FIGURE 11.—*Vertebrae and ribs of* Kotlassia prima
Natural size.

The succeeding six or seven vertebrae differ from the third by a somewhat more vigorous development of their neural arches, pre- and postzygapophyses, and bodies. The articular surfaces on the transverse processes keep their oval outlines, but their size from the fourth to the tenth vertebrae decreases gradually. The size of the hypocentrum remains unaltered (Figs. 8, 9).

The rib of the fourth vertebra has a strongly developed tubercle; the cervex of the rib is very long, the distal end is very much widened (Fig. 8).

The proximal ends of the ribs of the fifth, sixth, seventh, and eighth vertebrae differ but little from the same end of the rib of the fourth vertebra. The difference consists only in their tubercles becoming gradually smaller. A large process, with its apex directed upwards and backwards, gradually becomes differentiated on the

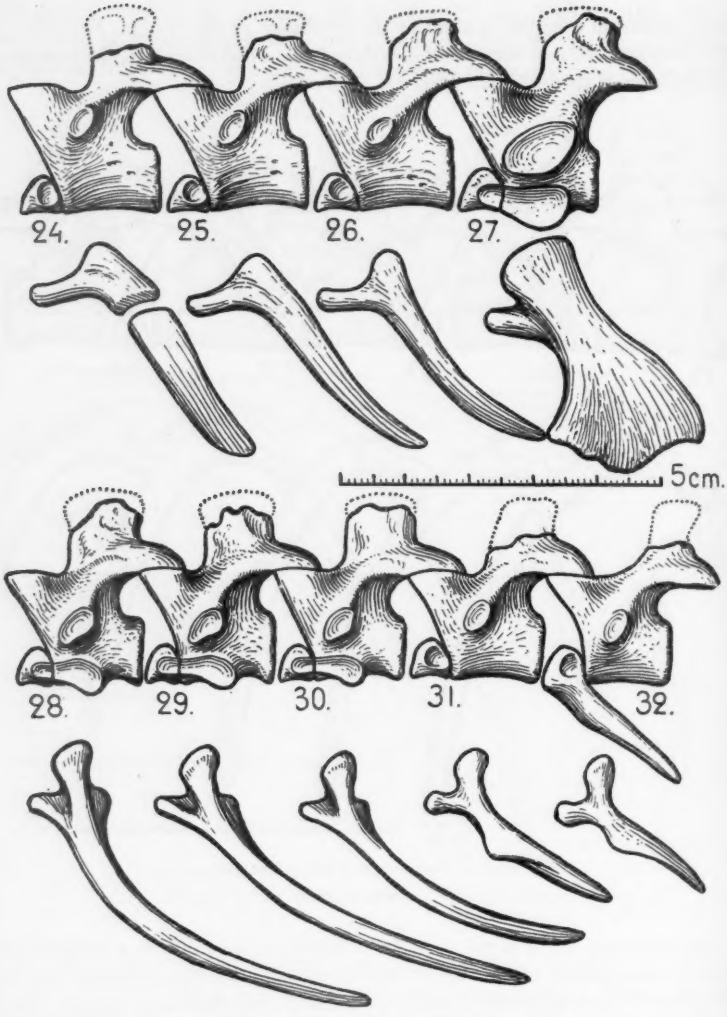

FIGURE 12.—*Vertebrae and ribs of* Kotlassia prima
Natural size.

widened distal extremities of the anterior ribs. It is developed best of all on the rib of the seventh vertebra (Fig. 8) and least of all on the rib of the ninth (Fig. 9).

The rib of the tenth vertebra has no expansion on its distal extremity.

From the eleventh vertebra to the twentieth, inclusively, there appears in the vertebral column of *K. prima* (No. 74) an interesting peculiarity. The centrum and

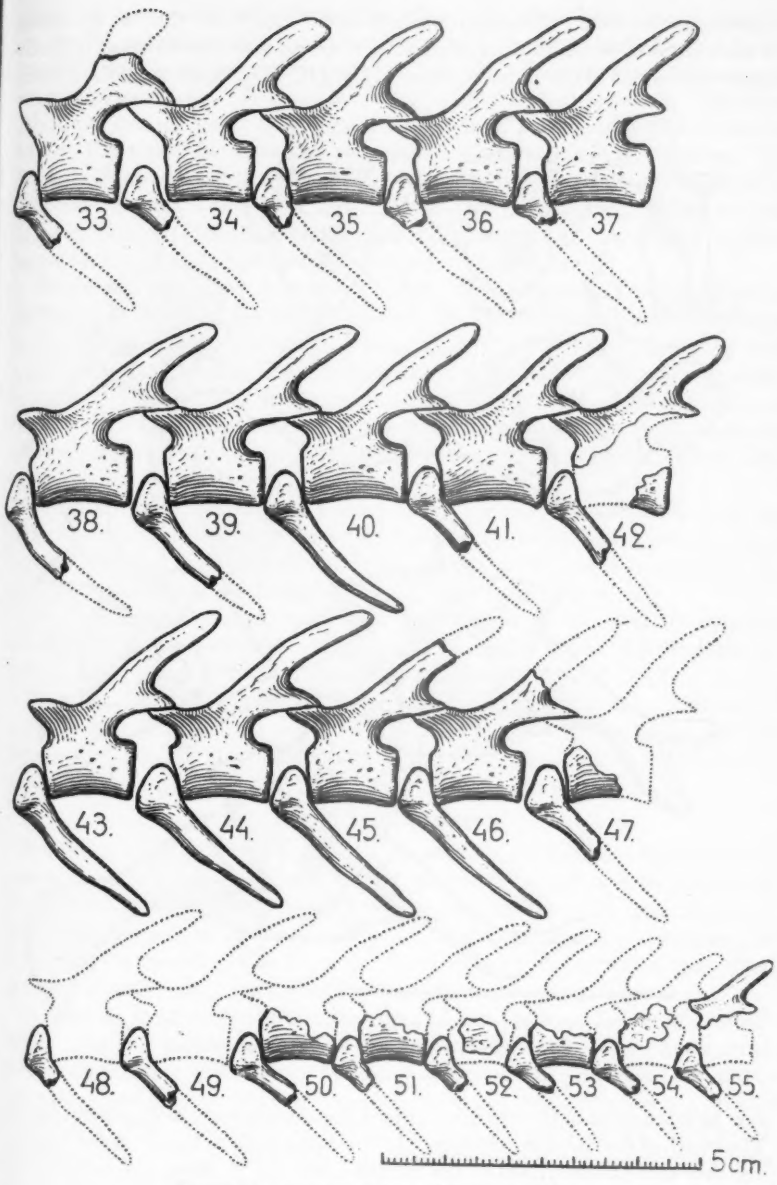

FIGURE 13.—*Vertebrae and ribs of* Kotlassia prima

Natural size.

hypocentrum of these vertebrae are entirely fused (Figs. 9, 10, 11). In *K. secunda*, as the additional preparation has shown, there are no such concrescences in the corresponding section of the vertebral column (Fig. 22). The character of the rib struc-

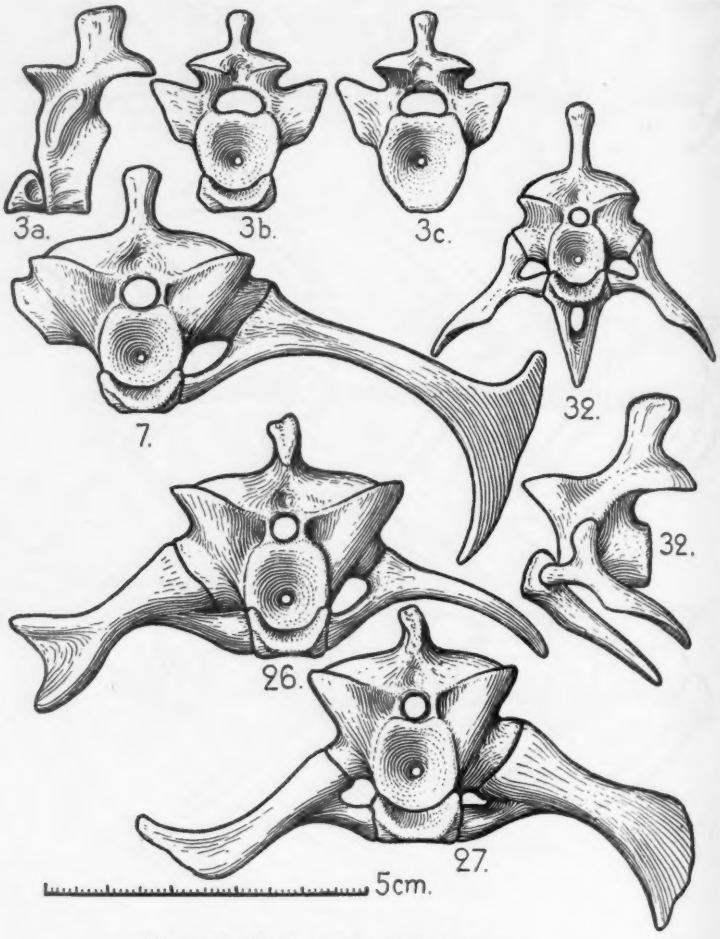

FIGURE 14.—*Vertebrae and ribs of* Kotlassia prima
Natural size.

ture in this section of the vertebral column remains constant in its general features. The length of the ribs alone changes. It increases a little to the rib of the thirteenth vertebra and gradually falls to the twenty-third. In *K. prima* the twentieth vertebra is the last where the centrum and hypocentrum are fused. Beginning at the twenty-first and up to the fifty-fifth vertebra, (the twenty-eighth caudal) the hypo-

centra are again independent bony structures. The ribs of the twenty-second to twenty-fifth vertebrae differ from those preceding by a certain widening and flattening of the middle section of their body (24, Fig. 12). The twenty-seventh vertebra is a typical sacral one. To it is joined a very powerful short sacral rib. The broad tubercle of this rib is articulated with a similar broad transverse process of the vertebra. The articulated surface on the transverse process has an oval outline. The long axis of the oval is inclined at an angle of about 45° to the longitudinal axis of the vertebral column. The head of the sacral rib is very broad and is joined not only to the hypocentrum but also with the centrum of the vertebra. As a result of this, a long articular area is formed on the lower edge of the vertebra.

The union of the capitulum of the rib with the hypocentrum and centrum is not an exclusive peculiarity of the sacral vertebra. The same type of articulation exists in the first three caudal vertebrae as well (28, 29, 30, Fig. 12). The ribs of these segments possess also very broad and flat heads each of which is united with a long articular area, situated on the lateral surface of the vertebra, and which embraces half the length of its centrum as well as its hypocentrum. The first three postsacral ribs are comparatively slender, bent and directed backwards by their acute distal ends.

The thirty-first vertebra bears a short rib, with a sharp crest situated on its anterior border. The small tubercle of this rib, as in the three preceding ribs, is turned dorsally; the head is of the same build as in the typical ribs and is articulated solely with the hypocentrum pertaining to the thirty-first vertebra (Fig. 12).

The rib of the thirty-second vertebra is the last of the series. It is like the rib of the thirty-first, but somewhat smaller. The peculiarity of the thirty-second vertebra is the presence of the first haemal arch on its hypocentrum (32, Figs. 12, 14).

The caudal vertebrae beginning with the sixth have no ribs (Fig. 13). Their neural spines become gradually narrower and higher, at the same time inclining strongly backwards. All their hypocentra bear well-developed haemal arches. The terminal caudal vertebrae (probably 10–15) are lacking in the skeleton.

APPENDICULAR SKELETON

The pectoral girdle (Figs. 15, 16) of *K. prima* (No. 74) has been sufficiently well preserved. The clavicle is rather strongly expanded and has a large prescapular process (Figs. 15, 16). The external surface of the clavicle in the place where the prescapular process goes off is covered by the same round cellular sculpture as is on the covering dermal bones of the skull. In passing to the ventral surface of the clavicle, the round cells become gradually transformed into elongated grooves.

On a rather considerable surface the edges of both the clavicles cover the interclavicle, situated on the medial line of the pectoral girdle. This bone has a pentagonal body with rounded angles. From one of these angles protrudes backwards a long parasternal process. At first it is rather wide, but then it narrows to a breadth of less than 1 cm. The ventral surface of the bone is covered by the usual round-celled sculpture, which changes into long grooves only on the periphery of the body and on the parasternal process.

Behind the prescapular process there is a well-developed cleithrum. Its upper

end is rather strongly widened; the lower end is covered by the prescapular process of the clavicle.

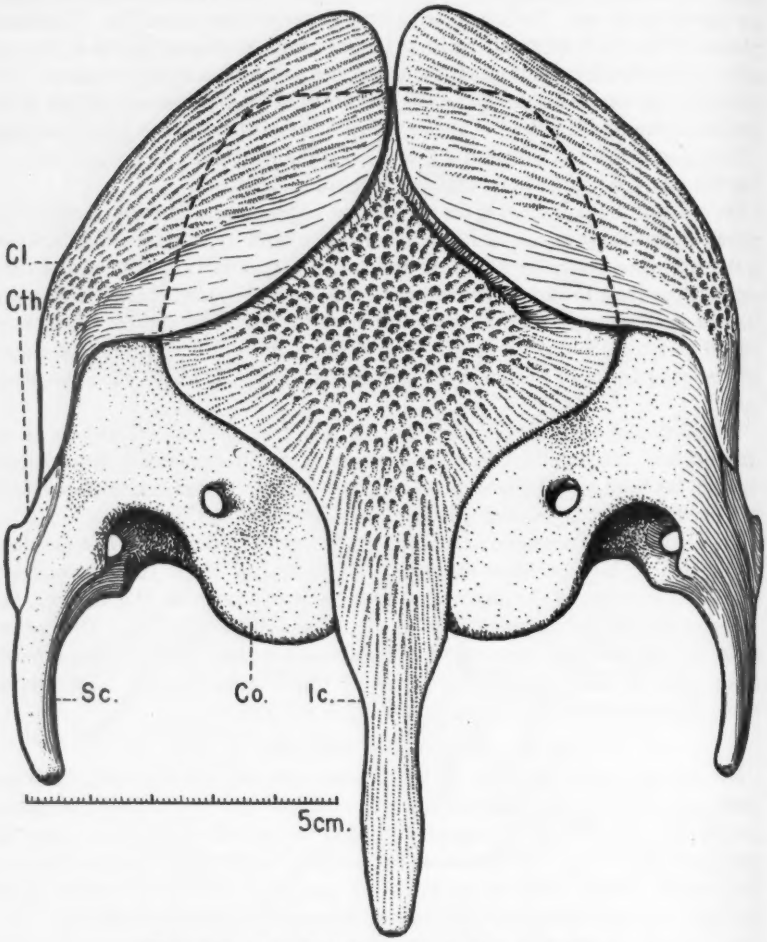

FIGURE 15.—*Pectoral girdle of* Kotlassia prima, *ventral view*
Cl—clavicle; Co—coronoid; Cth—cleithrum; Ic—interclavicle; Sc—scapula. Natural size.

To the cleithrum and the prescapular process of the clavicle attaches posteriorly a strongly ossified scapula with a rather large supraglenoid foramen (Fig. 15).

The coracoid in the pectoral girdle is completely ossified and forms a unit with the scapula. Not far from the cavitas glenoidalis, passed the supracoracoid foramen through the coracoid plate (Fig. 15).

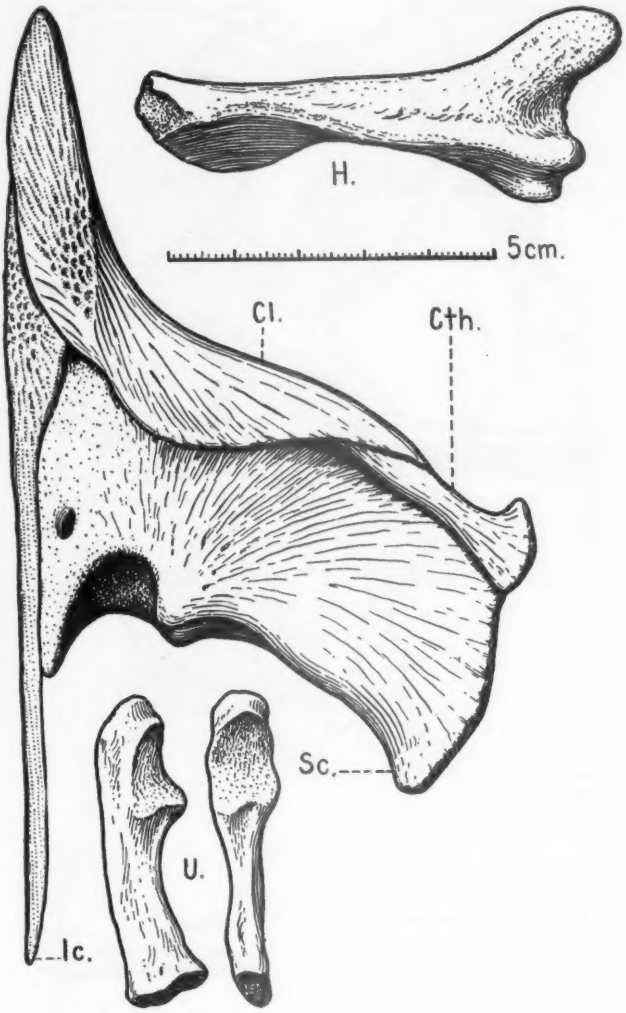

FIGURE 16.—*Pectoral girdle and limb of* Kotlassia prima

Cl—clavicle; Cth—cleithrum; Ic—interclavicle; H—left humerus, superior view; U—left ulna. Natural size.

The anterior extremity of *Kotlassia* has not been entirely preserved. The remains consist only of a humerus and ulna of the left side (Fig. 16). The humerus is a rather long bone with a well-developed deltopectoral crest and a comparatively wide distal end. The ulna is massive, flattened, with a well-formed and ossified olecranon. Radius, carpalia, metacarpalia, and phalanges are missing in the skeleton.

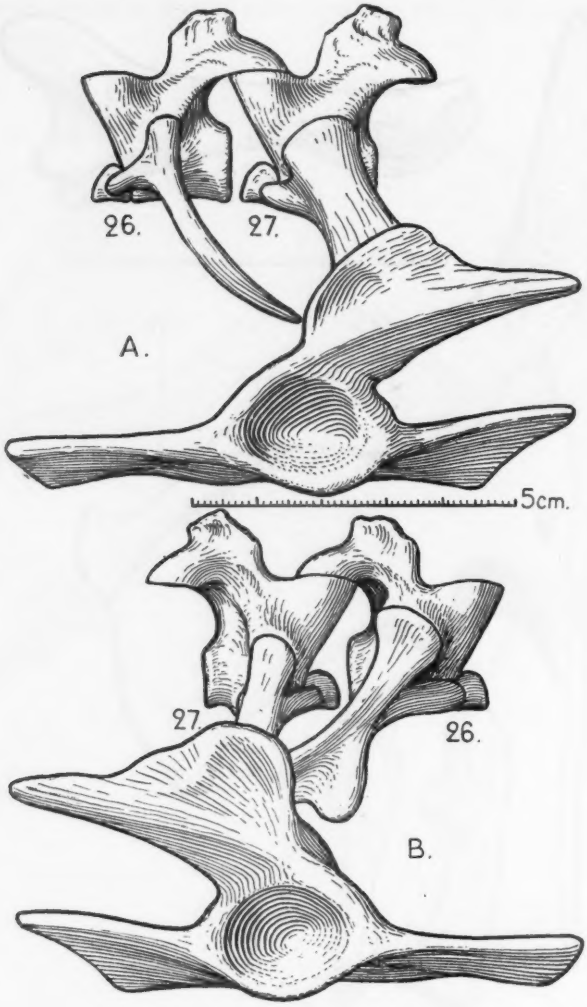

FIGURE 17.—*Pelvis of* Kotlassia prima, *lateral view*
Natural size.

It has already been mentioned that the twenty-seventh vertebra in *K. prima* (No. 74) is the sacral. An examination of the pelvic girdle on the left (Fig. 17A) shows that it is only this vertebra that bears a real sacral rib. On the left side, a normal rib of the presacral section of the body is joined to the twenty-sixth vertebra. However, the right rib of the same (twenty-sixth) vertebra is incomparably more developed than the left (Fig. 14) and is shaped like a real sacral rib. This is furnished

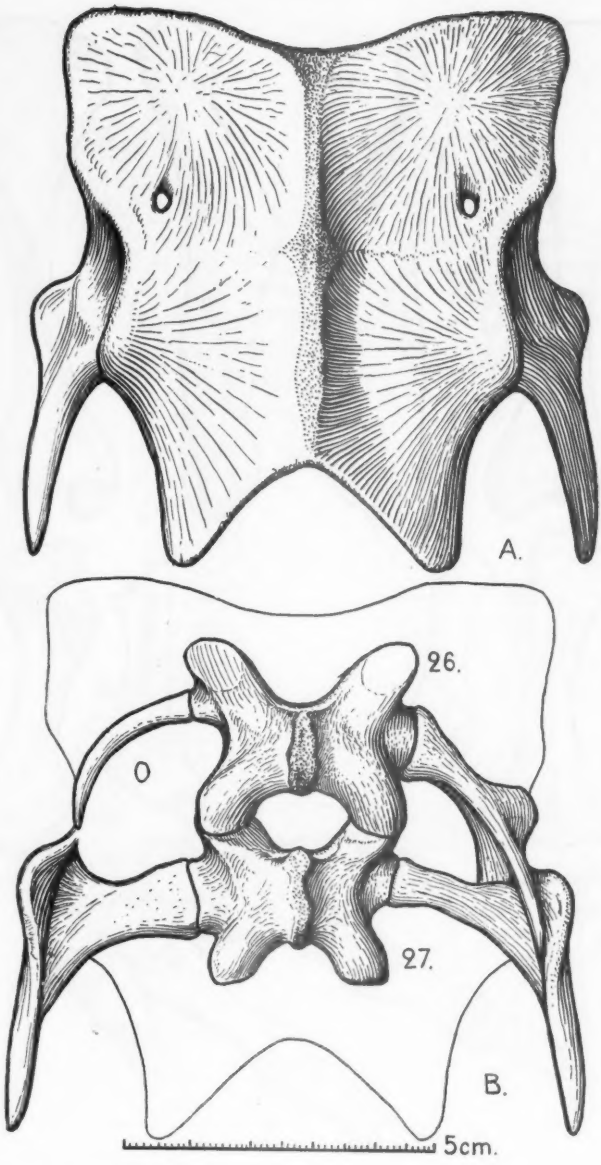

FIGURE 18.—*Pelvis of* Kotlassia prima, *ventral and dorsal views*

A—Ventral view; B—dorsal view. Natural size.

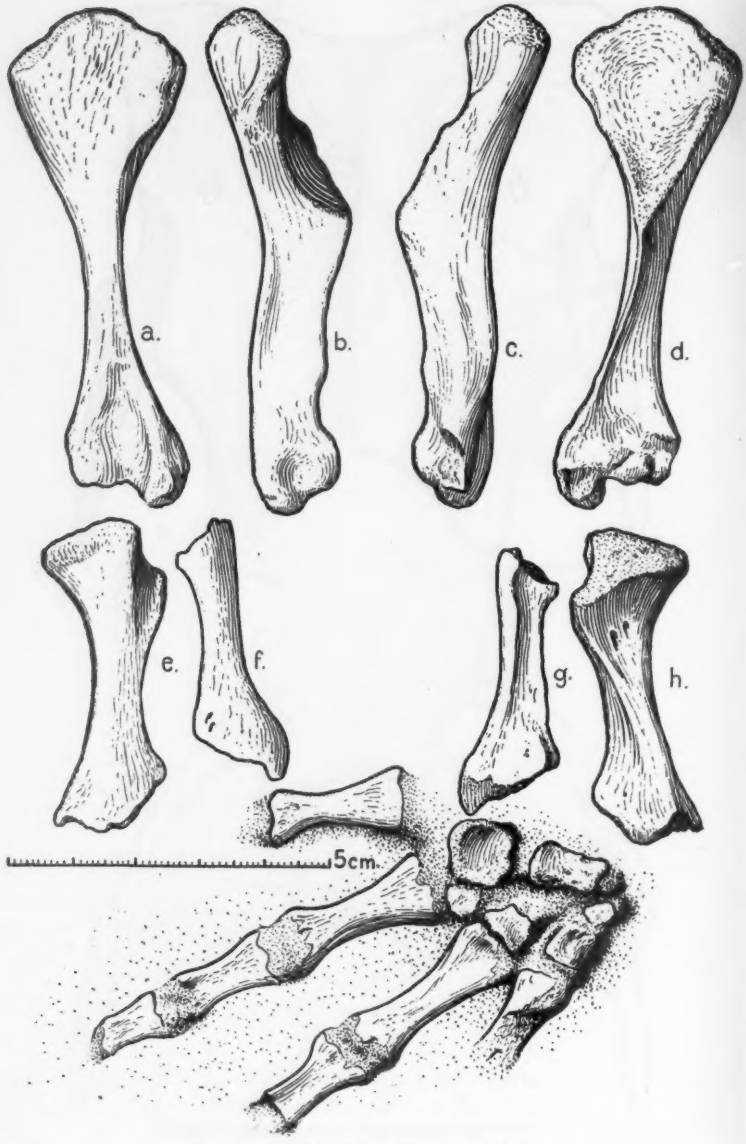

FIGURE 19.—*Pelvic limb of* Kotlassia prima

a, b, c, d,—left femur; e, f, g, h,—left tibia and fibula. Natural size.

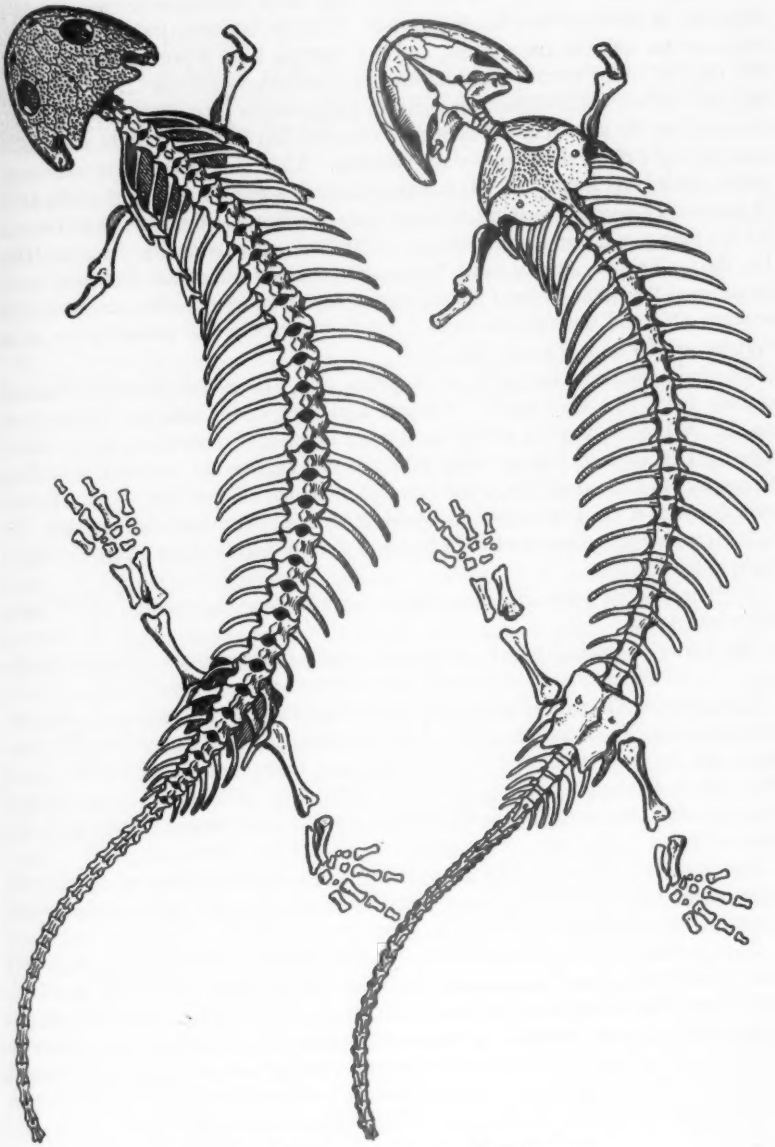

FIGURE 20.—*Restoration of the skeleton of* Kotlassia prima ×⅛.

with a large tubercle united to a correspondingly large transverse process; the distal extremity of the rib is very much widened. Thus, in the twenty-sixth vertebra the right and the left ribs are unequal, and the vertebra itself is not quite symmetrical (26, Fig. 14). The twenty-seventh vertebra bears two sacral ribs (27, Fig. 14), but they also are not quite similar, the right rib being weaker than the left; it is not only thinner than the left rib of the same (27) vertebra, but it is also weaker and smaller than the right rib of the twenty-sixth vertebra. The union of the vertebral column to the pelvis in *K. prima* (No. 74) is realized in such a manner that on the left one rib of the twenty-seventh vertebra is joined with the ilium, and on the right two ribs (of the twenty-sixth and twenty-seventh vertebrae) articulate with the ilium (Figs. 17, 18B). This peculiarity must be considered as an individual variation in the structure of the pelvic girdle of the skeleton of *Kotlassia*. In another specimen of the skeleton (No. 81) the pelvis is built quite symmetrically and possesses one sacral vertebra with a pair of sacral ribs.

The ilium, ischium, and pubis in *K. prima* (No. 74) are completely ossified and entirely fused with each other. A rather wide wing of the ilium is inclined backwards. Where it joins the sacral rib, it has a long sharp process caudally directed. The right ilium (Fig. 17B) is wider than the left, because the former is joined not to one but to two sacral ribs. Ischium and pubis have fused (Fig. 18A). Not far from the edge of the acetabulum, the pubis is pierced by a rather wide opening. On the ventral surface, along the line of fusion of the two halves of the pelvis, is a rather high, rough crest.

The bones of the pelvis in *K. secunda* (= *Karpinskiosaurus secundus* No. 81) have fused completely.

The posterior extremity in *K. prima* has been better preserved than the anterior one. In the skeleton No. 74 there are nearly all the bones of the left hind leg.

The femur of *Kotlassia* (Fig. 19) is a little longer than the humerus. Its proximal extremity is wide and flattened (Fig. 19a); the distal end is narrower than the proximal. On the lower surface of the bone there is a sharp crest (linea aspera, b, c, d, Fig. 19); the trochanter is feebly expressed (d, Fig. 19). The tibia is rather massive, especially in its upper half. On the anterior surface of its proximal extremity a high crest is visible (e, Fig. 19). The posterior surface of the bone is concave.

The fibula is much weaker than the tibia. On its anterior surface, along its lateral edge passes a somewhat high sharp crest; the anterior surface of the bone is quite smooth (f, g., Fig. 19).

The bones of the foot in *K. prima* (No. 74) have not been separated from the matrix and are seen only from their plantar side. Among the tarsal bones one can distinguish incompletely ossified fibulare, intermedium, a small fragment of tibiale, and three half-destroyed tarsalia. Of metatarsal bones only four have been preserved in the skeleton, and of phalanges but two on one toe, one on another. Owing to the lack of other skeletons with better-preserved extremities it is impossible to determine exactly the number of phalanges in each digit of *Kotlassia*.

RESTORATION

The total length of the skeleton of *K. prima* (No. 74) from the anterior edge of the skull to the last-preserved caudal vertebra is about 125 cm. Judging by the structure

of the skeleton (Fig. 20) the animal was long and narrow, with a long tail and comparatively strong extremities.

The length of the skeleton of *Kotlassia secunda* (*Karpinskiosaurus secundus*, No. 81) cannot be exactly determined, as a certain number of vertebrae are missing in the presacral section, and all but the first three caudal vertebrae.

ARMOR

The elements of the armor (Fig. 21) described by A. P. Hartmann-Weinberg (1935), preserved in the North Dvina collections of V. P. Amalitzky apparently really belong to *Kotlassia* as indicated by the structure of several ribs, which are included in a piece of matrix together with the armor, lying on its internal side. These ribs do not differ in any way from those of *K. prima*, belonging to the vertebrae from 10–11 to 20–21.

The preserved fragment of the armor is formed of several rows of bony plates, whose external surface is covered by a quite peculiar sculpture (Fig. 21A). On every element of the armor one can distinguish three longitudinal crests (Fig. 21B). One of these crests lies in the middle of the armorplate and is not equally well developed on all the elements. The other two parallel the first, almost halfway between the middle crest and the lateral edge of the plate. All three are joined by transverse bars. Analogous bars go likewise from the lateral crests to the edges of the bones.

The additional preparation of the armor fragment enabled us to establish that those elements which were situated on the median line, along the vertebral column, bear on their inner side rather long and wide lamellate processes (Fig. 21 C, D). These processes are perpendicular to the sculptured plates on a longitudinal plane. The other bony plates of the armor, situated on both sides of the median row, lack such processes.

The lamellate processes of the middle row of armor elements entered undoubtedly into the layer of conjunctive tissue between the bands of the epaxial musculature. Their ends probably reached the summits of the neural spines but did not penetrate between them. The distance between the neural spines of two neighboring vertebrae in all *Kotlassia* is less than the width of the processes of armor elements, favoring this opinion.

The ribs found with the fragment of armor, to judge by their structure, belong to those vertebrae where the hypocentra coalesce completely with the centra (eleventh to twentieth vertebrae). Very likely the presence of armor diminished the mobility of the section of the vertebral column situated immediately below, and this, in turn, caused the coalescence of the hypocentra. If so, the more or less resisting armor of *Kotlassia* was present only at the level of the eleventh to twenty-first vertebrae. At the front and at the back of this section, the armor was either totally absent or was formed of bony elements less closely drawn together.

DISCUSSION

SYNONYMY OF DESCRIBED SPECIES

V. P. Amalitzky, in describing two separate species of *Kotlassia*, considered, as the specific peculiarities of *Kotlassia prima*, somewhat greater dimensions of the skeleton, a more rounded head, a shallower otic notch, and strong ribs.

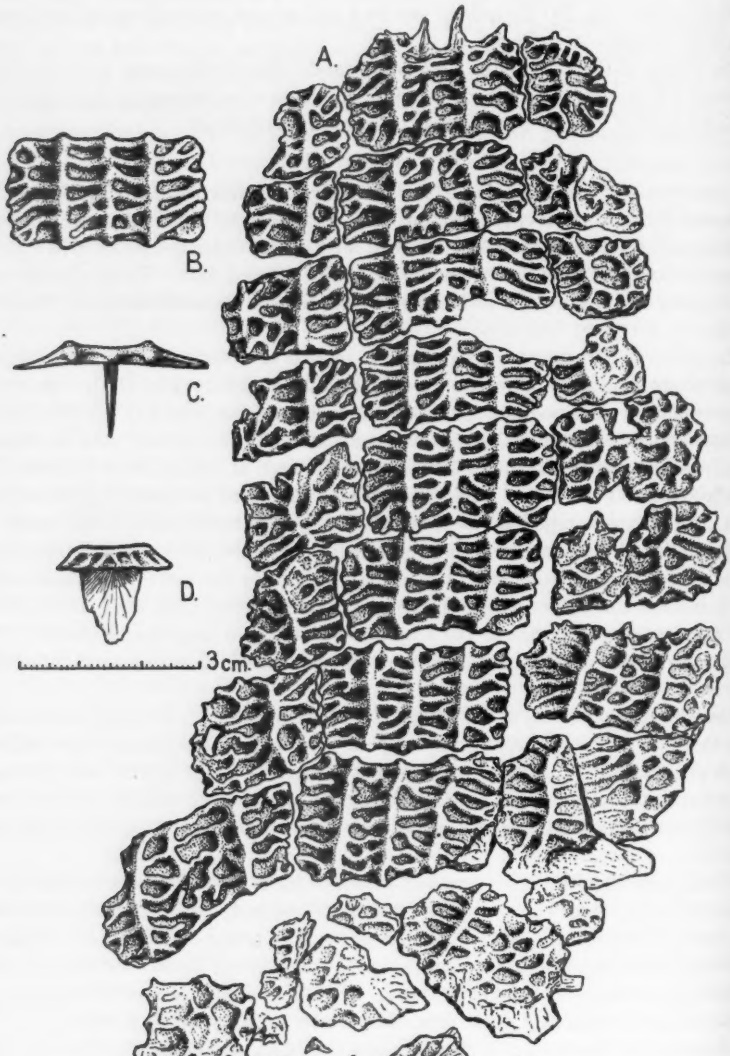

FIGURE 21.—*Armor of* Kotlassia prima

A—Armor of *Kotlassia prima*; B, C, D—the separate elements of the armor. Natural size.

The peculiarities of *Kotlassia secunda*, according to Amalitzky, are: large neural spine of the vertebra, comparatively short ribs, and wide cells on the dermal bones of the skull.

P. P. Sushkin found still greater differences between the two species of *Kotlassia*. In his opinion *Kotlassia secunda* (No. 81) differs so much from *K. prima* that it ought to be considered a separate genus. Therefore, Sushkin changed the name of *Kotlassia secunda* to *Karpinskiosaurus secundus*. As the most important distinctive peculiarities of *Karpinskiosaurus*, Sushkin considered the large neural spines, the absence of bony intercentra (= hypocentra) of the vertebrae, two sacral vertebrae, and a somewhat lesser widening of the distal ends of the first ribs.

A study of all the material of the North Dvina *Kotlassia*, plus additional preparation of the skeletons indicates that the differences pointed out by Amalitzky and Sushkin either do not exist or else have not the importance of specific characters.

Deformation of the skulls of *Kotlassia prima* and *K. secunda* (= *Karpinskiosaurus*) makes it impossible to speak with sufficient assurance of the differences in their general configuration. The writer believes that, if this deformation is evaluated, the skulls of all the *Kotlassia* specimens will differ only in size and not in shape. The bones in the region of the otic notch in *K. prima* have not been sufficiently preserved. This makes it impossible to compare the otic notches on the skulls of *K. prima* and *K. secunda*. In the author's opinion there is no need to seek for differences where their existence cannot be proved by material facts.

The sculpture on the dermal bones of *K. prima* is poorly preserved. However, the small bone fragments on which it exists seem to show that the sculpture of *K. prima* did not differ from that of *K. secunda*.

The neural spines of both skeletons cannot be compared, as they have only been preserved on certain vertebrae in the skeleton of *K. secunda* and have all been, except on the third vertebra, broken off and destroyed in the first preparation of *K. prima*. The dimensions of those fragments of neural spines which remain show that in *K. prima* they were not smaller than in *K. secunda*.

The distal extremities of the anterior ribs in *K. secunda* are really less broadened than in *K. prima*. However, this may, on serious grounds, be considered rather as a peculiarity of age than a specific difference, all the more so as the general dimensions of the *K. secunda* skeleton are smaller than those of *K. prima*.

Sushkin speaks of the presence of two sacral vertebrae in *K. secunda* remarking, however, that one of them is very feebly joined to the pelvis. Additional preparation has enabled us to establish that there is but one sacral vertebra in the skeleton of *K. secunda* (= *Karpinskiosaurus*) (Fig. 22). The ribs of the neighboring vertebrae have no contact with the ilium.

The statement of Sushkin regarding the absence in "*Karpinskiosaurus*" of bony intercentra (=hypocentra) in the first caudal vertebra cannot be understood at all, as they certainly exist on the sections of the vertebral column that have undergone preparation. In addition, they have been also successfully prepared where they had previously been covered by matrix.

The undoubted differences between the skeletons of *Kotlassia prima* (No. 74)

and *Kotlassia secunda* (=*Karpinskiosaurus*) (No. 81) are on the whole reduced to the following: the dimensions of the skull, the size of the vertebrae, the width of the distal extremities of the anterior ribs; the length of the other ribs in *K. prima*

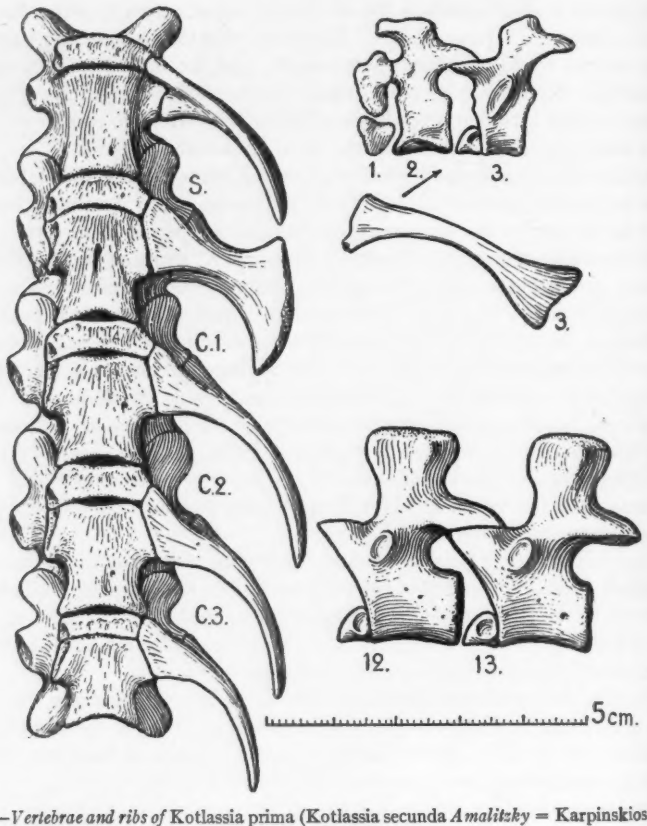

FIGURE 22.—*Vertebrae and ribs of* Kotlassia prima (Kotlassia secunda *Amalitzky* = Karpinskiosaurus secundus *Sushkin*)

Natural size.

is greater than in *K. secunda*; the hypocentra of the eleventh to twentieth vertebrae, coalesced in *K. prima* with the corresponding centra, remain free in *K. secunda*. This has been shown by the special additional preparation of skeleton No. 81 (Fig. 22).

All these differences must be considered only as age variations in the skeletons of the same form. It has not been possible to detect any specific differences in the two skeletons (No. 74 and No. 81) except those age differences pointed out.

The collection of Amalitzky contains a strongly deformed skull (No. 82). Ama-

litzky considered that it belonged to *K. secunda*. Sushkin, taking into account the somewhat smaller sized cells on its dermal bones, separated it into a distinct genus *Karpinskiosaurus*, naming the species *Karpinskiosaurus neglectus*.

A study of the dermal bones and of the changes in their ornamentation during the process of individual growth of the skull in the labyrinthodonts (Bystrow, 1935) has shown that, as the size of the bone increases, that of the cells does also. Keeping in mind the smaller size of the *Karpinskiosaurus neglectus* skeleton one can affirm with certainty that it is but a younger specimen of *Kotlassia prima*.

A complete analogy in many structural details of the skeletons, and the fact that there is only a small difference in their dimensions, leads to the conclusion that the collection of Amalitzky contains only one form of *Kotlassia*. It must be named *Kotlassia "prima"*, although the specific name "*prima*" is not a happily chosen one.

RELATIONSHIPS

The additional preparation of the skeletons of *Kotlassia* provided with sufficient accuracy many heretofore unknown details of its structure. The author's study of these details causes him to support Sushkin's opinion that the North Dvina *Kotlassia* is an amphibian.

First of all this appears from the structure of its skull. The skull roofing of *Kotlassia* is certainly stegocephalic. The same bone elements enter into its constitution which form the roofing of the skull in labyrinthodonts. In *Kotlassia* there is a well-developed intertemporal, which is not encountered as an independent bone in all labyrinthodonts. This bone can be seen, for instance, in *Loxomma*, *Palaeogyrinus* (Watson, 1926b), *Pelion lyelli*, *Leptophractus* (Romer, 1930), *Dendrerpeton*, *Platystegops* (Steen, 1930), and *Trimerorhachis* (Case, 1935).

The lacrimal in *Kotlassia* takes part in the formation of the edges of the external naris and orbit as in *Dvinosaurus* (Bystrow, 1938a), *Micropholis stowi* (Watson, 1913), *Palaeogyrinus decorus* (Watson, 1926a).

All the dermal bones of the *Kotlassia* skull are covered by a typical round-celled sculpture. In this respect it does not differ in any regard from the usual labyrinthodonts.

The furrows of the lateral line system on the skull of *Kotlassia* are lacking, but they also do not exist on the skulls of certain stegocephalians: *Platyops watsoni* (Efremov, 1933), *Diplocaulus*, *Diceratosaurus*, *Melosaurus*, and *Platyrhinops mordax* (Steen, 1930).

The stapes of *Kotlassia* is a little shorter than in the labyrinthodonts, but, as in them, it is directed at its tip to the center of the otic notch, over which the tympanic membrane was stretched. The otic notch itself is bounded by tabular, supratemporal, and squamosal, as in embolomerous stegocephalians (Efremov, 1933).

The twelfth pair of the cranio-cerebral nerves, without any doubt, had no special opening in the skull of *Kotlassia*.

The palate of *Kotlassia* is built as in *Loxomma*, *Baphetes*, *Pholidogaster* (Watson, 1926), and *Macromerion* (Romer, 1930).

The sphenethmoid, lying on a long processus cultriformis, does not differ in shape from the sphenethmoids which we find in *Wetlugasaurus* and *Zygosaurus*.

The lower jaw of *Kotlassia*, by the number and the disposition of the dermal bones of which it is formed, can only be considered as the mandible of a labyrinthodont.

The teeth according to their structure must also be recognized as the teeth of a labyrinthodont.

The same bones, including the cleithrum, form the pectoral girdle of *Kotlassia* as in the stegocephalians. The cleithrum has not been found in *Seymouria baylorensis*. The humerus in *Kotlassia* has no entepicondyloid foramen.

The pelvic bones, by means of the sacral ribs, are joined to one sacral vertebra, if one does not take into account the individual variation on the right side in the skeleton of *Kotlassia prima* (No. 74).

Thus, the structure of the skull, teeth, pectoral, and pelvic girdles of *Kotlassia prima* is that of a stegocephalian, and not some form of Reptilia. This stegocephalian, owing to numerous morphological peculiarities of its cranium, must apparently be recognized as very near the ancient stegocephalians of the Embolomeri type. However, the hypocentra in *Kotlassia prima* are reduced almost to the same degree as in *Labidosaurus hamatus* and *Captorhinus*, which makes its vertebrae very unlike those of the embolomere type.

The vertebrae of *Kotlassia* are certainly of the reptilian type. It does not, of course, follow that *Kotlassia* ought to be regarded as a direct ancestral form of Reptilia. Its Upper Permian age certainly does not allow us to make such a supposition. However, *Kotlassia* can be regarded as a relict form, reminding one by its organization of the amphibian ancestors of Reptilia. This, undoubtedly, constitutes the great phylogenetic interest of *Kotlassia*. In recognizing *Kotlassia prima* as an amphibian, we indicate that among the stegocephalians is a form with a structure of vertebrae that is incomparably nearer those of Reptilia than those of Embolomeri.

The question as to how the vertebra of *Kotlassia* has developed, and through what stages it passed in the process of this development, cannot be decided at present, owing to the lack of necessary materials. However, the initial point of its development may have been an embolomerous vertebra of a structure like that of *Dendrerpeton* (Steen, 1934). Out of this vertebra by a gradual reduction of basiventrale (=intercentrum = hypocentrum) a small hypocentrum could have been formed, while a strong development of interventrale yielded the fundamental mass, the centrum, to form the body of the vertebra.

ENVIRONMENT

The distribution area of *Kotlassia* was distinguished by marked aridity. The fauna of North Dvina, where *Kotlassia* has been found, consisted of terrestrial theromorph Reptilia (*Inostrancevia*, *Dicynodon*, and others) and of cotylosaurs (*Pareiasauria*) inhabiting the banks of rivers flowing through a dry steppe.

Sushkin (1936) and Efremov (1933) considered the Upper Permian climate of the northern region of the European part of U. S. S. R. most unfavorable for the existence of stegocephalians, as they are more closely connected with water basins than Reptilia. This explains the absence in the North Dvina fauna of typical labyrinthodonts other than *Dvinosaurus*; this form could persist in unfavorable climatic conditions because of its life-long retention of gill breathing.

However, if *Kotlassia* is considered an amphibian, *Dvinosaurus* is not the only representative of the stegocephalians in the North Dvina fauna. *Kotlassia*, without doubt, was the second stegocephalian in this fauna.

If a comparison be made between *Dvinosaurus*, which has retained gill breathing, and *Kotlassia*, the "terrestrial" organization of the latter becomes evident.

The furrows of the lateral line systems on the dermal bones of the *Dvinosaurus* skull are very well developed, while on the skull of *Kotlassia* they are totally absent. This, evidently, shows that the adult *Kotlassia* was not aquatic.

The pectoral and pelvic girdles and the extremities of *Kotlassia* are relatively stronger than the same parts in the skeleton of *Dvinosaurus*. The general character of the skeletal structure in *Kotlassia* reminds one very much of the Reptilia. Nevertheless, with all its "terrestrial" nature, *Kotlassia*, like a true stegocephalian, certainly needed water and passed through all the larval stages of its development in the same water basins, which were inhabited by such a neotenic form of labyrinthodont as *Dvinosaurus*.

COMPARISON WITH *SEYMOURIA*

It is somewhat difficult to compare *Kotlassia prima* and *Seymouria baylorensis*, as the latter has not yet undergone sufficient preparation and cannot be considered as completely studied, in spite of the fact that the materials in the United States museums are sufficiently rich.[2]

The skull, anterior section of vertebral column, and a fragment of the pectoral girdle of *Seymouria baylorensis*, described by Broili (1904a), were prepared so badly that one cannot obtain a clear idea of these parts.

The figure of the skeleton of *Seymouria*, widely current in the literature, first published by Williston (1911), and the plaster reconstruction made, following Williston, by Romer, cannot serve as materials for comparing *Kotlassia* and *Seymouria*. In these reconstructions there are too many skeleton bones which one cannot with certainty attribute to *Seymouria*.

As yet no complete skeleton of *Seymouria* has been described, although Romer (1928) mentions that in the Chicago Museum there is an almost entire skeleton, not yet prepared, included in very hard rock (No. 663, Walker Museum, University of Chicago).[3]

The greatest amount of morphological data on the skeleton of *Seymouria* are given by Watson (1918). Nevertheless, these data likewise allow us to compare the two forms only in their most general features.

Watson (1918) found in the skull of *Seymouria* small bony elements on both sides of the base of the processus cultriformis (between this and the pterygoid). Broom (1922) looked upon these formations as ossified "mesopterygoids", which he found in the form of cartilaginous elements in studying the skull of the embryo of *Agama hispida*. However, the bony elements found by Watson probably are the same basipterygoid processes of the basisphenoid as in *Kotlassia*, and not independent formations.

Watson (1918) points out that in *Seymouria* the fenestra ovalis is bounded by four

[2] *Cf.* Postcript.

[3] Actually, this is the skeleton described by Williston.

bones, namely, exoccipital, paroccipital, pro-otic, and basisphenoid. In *Kotlassia* only the paroccipital and pro-otic participate in the formation of this opening.

The lower jaw of *Seymouria* has been very inadequately studied. Broili (1904a) could not describe it minutely with the material at his disposal. Watson (1918) described only a small fragment of the mandible. The drawing, extant in literature, of the lower jaw of *Seymouria baylorensis* (Haughton, 1929–1930) presents only a reconstruction, made after a fragment published by Watson.

On the coronoid of *Seymouria*, Watson found one tooth. All endeavors to find traces of teeth on any of the coronoids of *Kotlassia* have yielded no results.

The structure of the teeth in *Seymouria* has been studied only by Broili (1927). He had at his disposal a cross section of a small part of one tooth. Broili turned his attention to the fanlike disposition of the dentinal canals in different sections of the tooth's wall. This enabled him to conclude that the tooth of *Seymouria* has a folded wall. Broili recognized that such a structure of the tooth distinguishes *Seymouria* from the representatives of the Cotylosauria, in particular from *Labidosaurus*, and draws it nearer to the labyrinthodonts. Nevertheless, he supports Watson, who considered *Seymouria* a reptile of the order Cotylosauria, and does not share the point of view of Sushkin and Broom, who see in *Seymouria* an amphibian.

According to the data of Broili, the teeth of *Seymouria* and *Kotlassia* are of analogous structure.

The neural spines of the vertebrae in *Seymouria baylorensis* are considerably smaller than in *Kotlassia prima*.

The type of union of the sacral rib with the corresponding vertabra was the same in *Seymouria baylorensis* and in *Kotlassia prima*. This is shown by the structure of the sacral vertebra sent to the Paleontological Institute of the Academy of Sciences of U.S.S.R. by Romer.

Watson had described (1918) a two-headed rib, on the first vertebra in *Seymouria*, and fourth and fifth caudal ribs. *Kotlassia* in this respect differs from *Seymouria*, as the first rib in its skeleton belongs to the third vertebra, whereas in the caudal section of the vertebral column the last rib is on the fourth vertebra.

The cleithrum is absent in the pectoral girdle of *Seymouria*; it is well developed in *Kotlassia*. On the contrary, the entepicondyloid foramen on the humerus of *Seymouria* is absent in *Kotlassia*.

No elements of a dorsal armor have been found in *Seymouria baylorensis*.

The existing data on the morphological peculiarities of *Seymouria baylorensis* are quite insufficient for proving the degree in which this form is similar to *Kotlassia prima*. These data are also insufficient to decide the question of the systematic position of *Seymouria*.

The absence of a detailed description of the North American Lower Permian *Seymouria* makes a decision as to its relation to the North Dvina Upper Permian *Kotlassia prima* exceedingly difficult.

One can accept as certain that *Kotlassia prima* is an amphibian, but one cannot be sure that *Kotlassia* belongs to the family Seymouridae. Only a new and detailed study of the North American materials on Seymouriamorpha could decide the question: Are *Seymouria baylorensis* and *Kotlassia prima* closely related forms?

The possibility is not excluded that *Seymouria baylorensis* represents an amphibian of the same kind as *Kotlassia*.

POSTSCRIPT

Sensing perhaps the shape of future events, Dr. J. A. Efremov, Curator of the Palaeontological Institute of the Academy of Sciences of the U.S.S.R., sent the above manuscript to me a short time before the German attack on Russia in 1941. He asked merely that the manuscript be kept in safety, although expressing the hope that the work might be considered for publication here. Doctor Bystrow's paper is a valuable contribution to our knowledge of an extremely important group of tetrapods.

It is unfortunate that Doctor White's detailed study of *Seymouria*[4] had not reached Doctor Bystrow before his manuscript was finished, for this remedies the lack of adequate description of the American material of which he justly complains. A parallel reading of the two papers is of interest. The structural situation is in many respects similar in the two types concerned; both authors agree that the Seymouriamorpha are morphologically close to the boundary between amphibians and reptiles, although White finally decides that the verdict is for assignment to the Reptilia, Bystrow to the Amphibia.

A. S. ROMER.

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c/o A. S. Romer, Museum of Comparative Zoology, Cambridge, Massachusetts.

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NOTE

THE MAP IS CONTOURED IN BOUGUER GRAVITY ANOMALY VALUES. ALL ANOMALIES ARE NEGATIVE AND ARE PLOTTED IN MILLIGALS. GRAVITY STATION LOCATIONS ARE INDICATED BY CIRCLES WITH THE NUMBER OF THE STATION NORTH OF THE CIRCLE, AND THE BOUGUER ANOMALY VALUE SOUTH OF IT.

Surveyed in 1941 by J. B. HERSEY

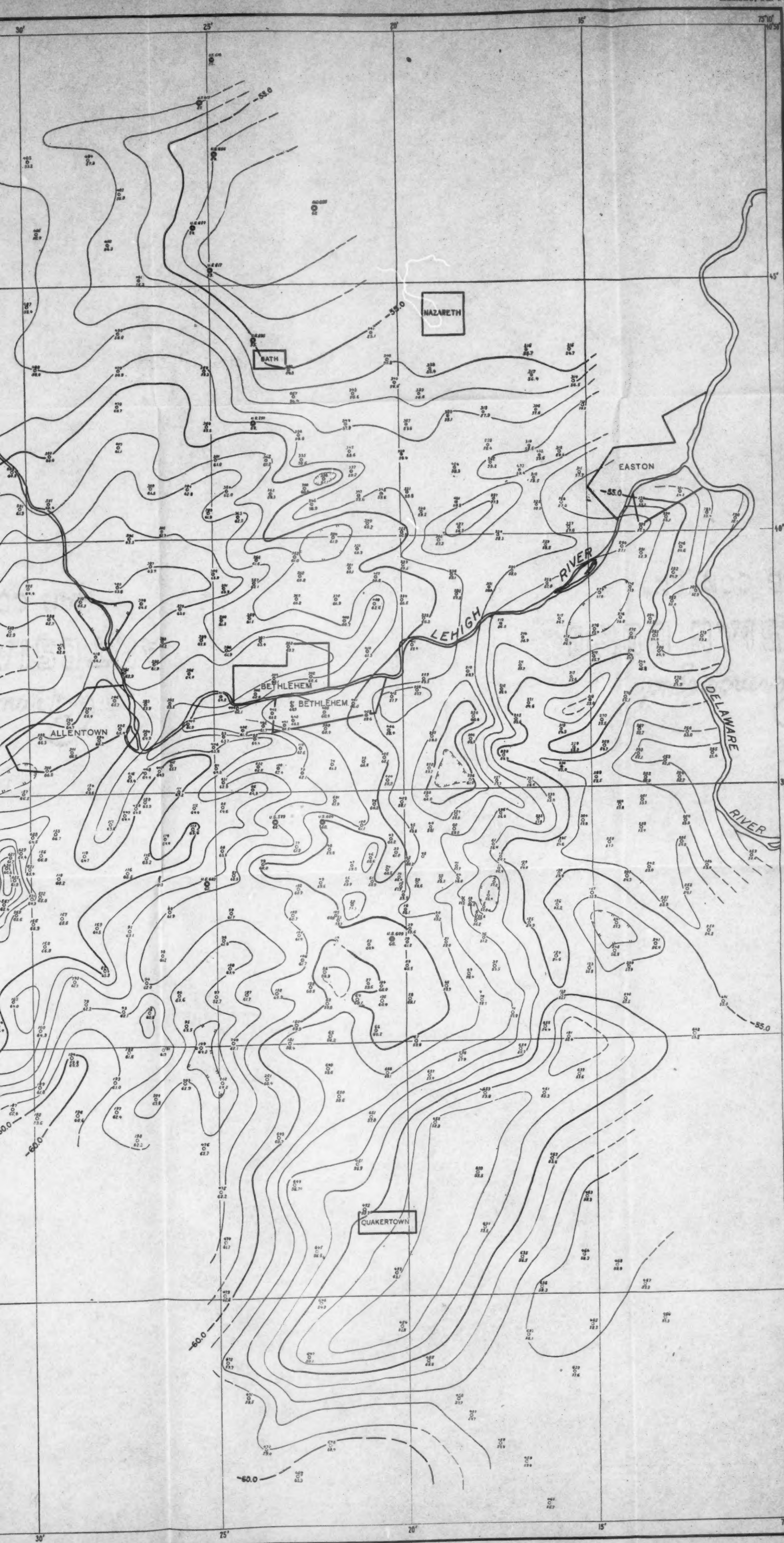
SCALE

1 ½ 0 1 2 3 4 Miles

1 ½ 0 1 2 3 4 5 Kilometers

Contour interval 1 Milligal

GRAVITY MAP OF CENTRAL-EASTERN



RAL-EASTERN PENNSYLVANIA

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GRAVITY INVESTIGATION OF CENTRAL-EASTERN PENNSYLVANIA

BY J. B. HERSEY

CONTENTS

	Page
Abstract	418
Introduction	418
Acknowledgments	419
Field procedure and gravity reductions	419
General statement	419
The gravimeter	419
Field procedure	419
Conduct of the gravity survey	420
Calibration of the gravimeter	420
Observed gravity computations	421
Accuracy of observed gravity values	423
Locations and elevations of gravity stations	423
Computation of the Bouguer anomaly	424
General considerations	424
Bouguer anomaly computations	426
Assumed density for the Bouguer correction	426
Terrain corrections	426
Results	427
Accuracy of results	429
Geologic interpretation of gravity anomalies in central-eastern Pennsylvania	429
General statement	429
Geology of central-eastern Pennsylvania	429
Stratigraphy	429
Physiography	429
Geologic structure	430
Density measurements	432
Interpretation of gravity anomalies	433
General considerations and methods	433
Anomalies due to simple forms	434
Depth estimation	434
Detailed analysis of mass anomalies	436
Interpretation of the results of this survey	437
Regional anomaly	437
Density contrasts	437
Local features	437
Individual local anomalies	438
Conclusions	443
Suggestions for future work	444
References cited	444

ILLUSTRATIONS

Figure	Page
1. Drift curve on June 18, 1941	421
2. Gravimeter calibration curve	422
3. Hexenkopf gravity map	428
4. Section of a spherical or cylindrical mass anomaly	434
5. Mass anomaly due to vertical faulting—a semi-infinite horizontal slab	435
6. Gravity anomalies over a sphere and an infinite horizontal cylinder	435
7. Gravity anomaly over a vertical fault	435
8. Gravity anomaly and section of the Triassic intrusive at Quakertown	440
9. Gravity anomaly across a syncline	443
Plate	**Facing page**
1. Gravity map of central-eastern Pennsylvania	417
2. Salisbury Church gravity map	430
3. Camels Hump-Pinetop gravity map	432
4. Areal geology of central-eastern Pennsylvania	444

ABSTRACT

In 1940 and 1941, 712 gravity stations were established in central-eastern Pennsylvania over a region centering about Allentown and Bethlehem. Geologically the area extends from the Triassic Lowlands to the Silurian of Kittatinny Mountain. Bouguer anomalies were computed, and from them gravity contour maps were made of the whole area and certain localities in it. Anomaly variation over the area is found to be both regional and local, the local anomalies correlating in shape with structural trends, and in several cases coinciding geographically with outcropping density contrasts that reasonably account for the gravity anomaly. Rough analysis is made of some of the structures, indicating the possible shape and size of mass anomalies that could produce the gravity anomalies found. Considerable overlapping of local effects makes detailed analysis difficult, but it is believed that the continued use of the gravimetric method in this region will be useful in studying its geologic and structural problems.

INTRODUCTION

In the spring of 1935 the United States Coast and Geodetic Survey commenced a program of gravity measurements along a profile in a north-south line through Bethlehem, Pennsylvania. This profile is about 60 miles long, has about 50 gravity stations, and was laid out to crosscut some especially interesting geologic structures. Although several stations were offset from the main profile to determine the anomaly variation along a strip rather than a single profile, the data proved to be inconclusive because an insufficient area was covered by an equally insufficient number of stations. This could not have been predicted prior to the survey, but in the meantime geologists in this part of the Appalachians have been interested in information to be gained from a sufficiently detailed gravity survey.

In the fall of 1940 a gravimeter of modern design and high precision became available to the writer through loan to the American Geophysical Union by the Humble Oil Company. With this instrument 71 stations were established near the Paleozoic-Triassic contact at Springtown, Pennsylvania, in the fall of 1940. During the spring of 1941, 641 additional stations were established covering a large part of Northampton and Lehigh counties and parts of Bucks and Berks counties, Pennsylvania. This report gives a detailed account of this gravity survey and discusses the significance of the gravity results in their relation to geologic problems in central-eastern Pennsylvania.

ACKNOWLEDGMENTS

Permission for the use of the gravimeter was granted by the American Geophysical Union, and Dr. G. P. Woollard very kindly loaned a special gravity field truck. Road profiles were obtained from the State Highway Department District Office at Allentown, and the Department of Civil Engineering at Lehigh loaned surveying instruments on several occasions. All these were vital to the success of the field work and are hereby gratefully acknowledged.

Many thanks are due the members of the staffs of the Physics and Geology departments at Lehigh for their interest and encouragement in this project and to several fellow students for their kind assistance in routine phases of the field work.

FIELD PROCEDURE AND GRAVITY REDUCTIONS

GENERAL STATEMENT

This section presents the details of the field procedure, including description of the instruments and methods used. The latter half of the section describes, with examples, the details of making the gravity reductions for use in geologic interpretation and outlines the general scope and character of the results.

THE GRAVIMETER

The gravimeter was designed and constructed by the Humble Oil Company. Like several others now in use it is entirely mechanical in operation and adjustment. Essentially, a relatively stiff spring counteracts most of the torque set up by the force of gravity on the heavy mass of the instrument. The remainder is balanced by a delicate, adjustable spring, controlled by a threaded shaft with a precision graduated counter on its head. The readings of this counter are observed as the arbitrary scale divisions to be calibrated in terms of gravity difference. The instrument is adjusted as a null reading instrument and is observed through an optical system that magnifies the mechanical motion.

The instrument must be carefully leveled and, in addition, is designed to compensate for small errors in leveling. Temperature effects are held to a minimum by housing the instrument in an insulating case and thermostating the internal temperature to within a few hundredths of a degree Centigrade. This control is adequate, and no temperature correction is necessary.

Controls for adjustment and operation are mounted on a panel on top of the case. Within the truck the instrument is mounted for observation on a tripod, the legs of which can be let down through the floor and clamped to be free of the truck. For transportation the instrument is mounted on a platform supported by the inflated inner tube of an airplane landing wheel and is supported laterally by four lengths of rubber tubing hooked to its four upper corners and the side walls of the truck.

FIELD PROCEDURE

In a typical field observation the truck is parked, the tripod is set up and roughly leveled, and the instrument is placed on the tripod and carefully leveled. The instrument is then released from its clamps and adjusted to the null position. The arbitrary scale-division readings are recorded, the final reading for a location being the one which repeats itself upon at least two settings to the null position. If there is a

noticeable falling-off level during the observation it is advisable to obtain more that two such checks when the observer is sure that the instrument is level.

The time of the gravity reading and the temperature inside the instrument are recorded, the station is described, and the terrain near the station is estimated. Some operators conduct the terrain survey somewhat elaborately with a separate party of surveyors. However, as pointed out in the section on terrain corrections, the locations were especially chosen to minimize terrain effects near the station, and therefore a detailed and precise survey of near-by terrain was not justified.

CONDUCT OF THE GRAVITY SURVEY

Since this is a relative method any single reading at one location is without significance. In order to obtain accurate gravity differences between a series of locations occupied in a single day's operation it is necessary to take readings at frequent intervals at a base station where the value of gravity is known. In the course of a single day the repeat readings at the base will not be identical because of tidal effects and instrumental changes. These effects constitute the drift of the instrument and must be corrected for in all the field observations of the day.

The drift correction is illustrated by the following data taken on June 18, 1941. The essential data and drift correction calculations are given in Table 1. The drift curve plotted from these data is shown in Figure 1. Field base station # 244 was used, and its assumed gravity value for the day is 587.0 scale divisions. It is assumed that the drift of the instrument was linear between successive observations at the base. This assumption is justified if base readings are repeated at least every 2 hours. At the end of the day the drift correction for each new station is read from the drift curve. From the corrected gravity values (Table 1) the proper gravity difference in scale divisions between the base and the other field stations occupied is computed.

CALIBRATION OF THE GRAVIMETER

The procedure outlined gives comparative values in terms of arbitrary scale divisions. In order to compute values in absolute units it is necessary to calibrate the instrument. Therefore, the first operation of a gravity survey is to occupy as many stations as possible where gravity is accurately known in order to convert the scale divisions into absolute units. In this region there are many U. S. Coast and Geodetic Survey stations. Reoccupations of many of these have shown that the instrument responds linearly to gravity changes over the range encountered in this region. The scale factor is 0.375 milligals per scale division. Table 2 gives the corrected data for calibration taken at several U. S. Coast and Geodetic Survey stations. Figure 2 is a plot of the arbitrary scale divisions against corresponding absolute gravity values in gals.

The calibration was checked at frequent intervals throughout the survey by means of a small counterweight placed on the moving arm of the gravimeter. There was no significant variation in the deflection caused by adding this counterweight at any time throughout the survey. From this it is concluded that there was no significant variation in the scale factor. As an additional check several reoccupations were made at the base stations. The results corroborate the above conclusion, and therefore the value 0.375 milligals per scale division has been used throughout the survey.

TABLE 1.—*Drift correction*

Station number	Time of observation	Uncorr. gravity value	Drift corr. (Values in arbitrary scale divisions)	Corrected gravity value
Base (244)	2:42	584.8	2.2	587.0
573	2:50	573.7	2.0	575.7
574	3:00	569.5	1.7	571.2
575	3:10	549.8	1.3	551.1
576	3:20	534.8	1.0	535.8
577	3:30	541.3	0.7	542.0
578	3:45	559.2	0.3	559.5
579	4:00	573.3	−0.2	573.1
580	4:15	588.2	−0.6	587.6
581	4:25	581.3	−0.9	580.4
Base (244)	4:35	588.2	−1.2	587.0
582	4:45	573.3	−1.1	572.2
583	4:55	557.7	−1.0	556.7
584	5:04	544.8	−0.9	543.9
585	5:18	566.3	−0.8	565.5
586	5:34	587.7	−0.6	587.1
587	5:44	591.0	−0.5	590.5
588	5:56	588.8	−0.4	588.4
Base (244)	6:12	587.2	−0.2	587.0

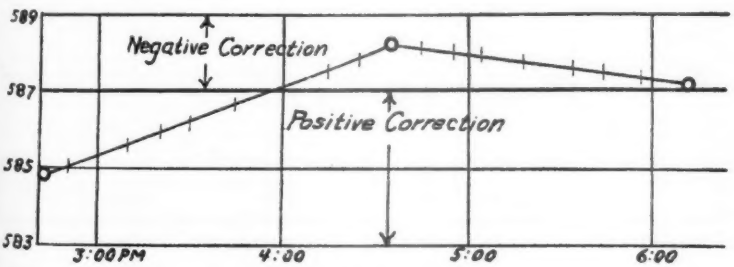


FIGURE 1.—*Drift curve on June 18, 1941*

OBSERVED GRAVITY COMPUTATIONS

The first major step in a gravity survey is the measurement of "Observed gravity" at the various stations. When the absolute value of the base station is known, the absolute value at any field station can be computed if the gravimeter has been calibrated. It is common practice to assume a value for the base station, and other stations are computed relative to the assumed value. In this survey the U. S. Coast and Geodetic station # 589 ("Lehigh University") was used as the primary

base. Its gravity value is published as 980.155 gals, with an accuracy of ±0.001 gals. Accordingly the value 980.1550 gals is assumed to be the correct value for this station, and all the stations of this survey are computed relative to it.

TABLE 2.—*Corrected data for calibration*

U. S. C. & G. S. gravity station		Absolute gravity (gals)	Corr. gravimeter reading (Scale divisions)
Number	Name		
589	Lehigh University	980.155	544.0
596	Bath	980.162	561.2
602	Friedensville	980.145	519.5
608	Center Valley	980.138	496.7
609	Amey	980.109	418.3
613	Green Pond	980.158	553.0
614	Kleintop	980.162	558.0
616	Hollo	980.167	571.0

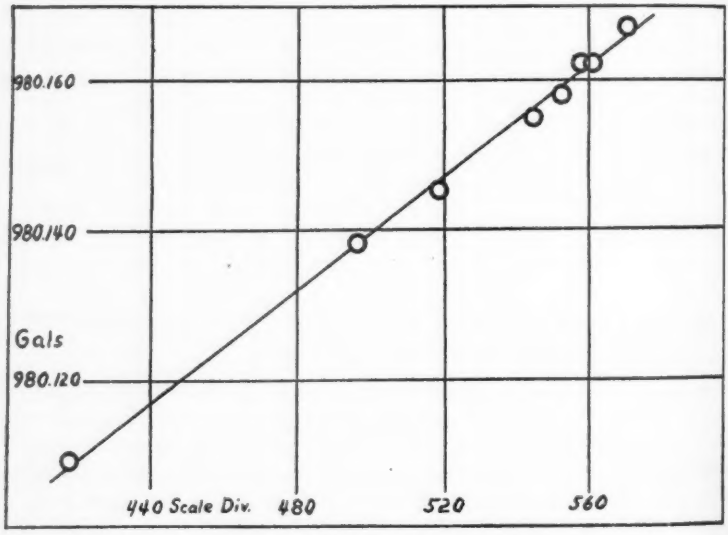


FIGURE 2.—*Gravimeter calibration curve*

The computation of observed gravity at stations occupied in one day is illustrated in Table 3. The corrected gravity values in scale divisions (second column) have been corrected for drift as observed at the base station. These values are multiplied by the scale factor, 0.000375 gals per scale division (third column). Gravity at the base station (*244 in this case) is already known. Therefore, if a correction be added to the third column to give the proper value at the base station, the same correction added to the third column for the other stations will give their value of

gravity to the same accuracy as the base. In the case of ✱244 gravity is 980.1611 gals. Therefore, the "correction to base" (fourth column) is:

$$980.1611 - .2201 = 979.9410.$$

Gravity at station 573, for example, is given by:

$$.2159 + 979.9410 = 980.1569 \text{ gals.}$$

TABLE 3.—*Computation of observed gravity*

Station number	Corr. gravity value (Scale div.)	Corr. gravity value × Scale factor (gals)	Correction to base (gals)	Observed gravity (gals)
Base (244)	587.0	.2201	979.9410	980.1611
573	575.7	.2159	"	980.1569
574	571.2	.2142	"	.1552
575	551.1	.2067	"	.1477
576	535.8	.2009	"	.1419
577	542.0	.2032	"	.1442
578	559.5	.2098	"	980.1508
579	573.1	.2149	"	.1559
580	587.6	.2203	"	.1613
581	580.4	.2176	"	.1586
582	572.2	.2146	"	.1556
583	556.7	.2088	"	980.1498
584	543.9	.2040	"	.1450
585	565.5	.2121	"	.1531
586	587.1	.2202	"	.1612
587	590.5	.2214	"	.1624
588	588.4	.2206	"	980.1616

ACCURACY OF OBSERVED GRAVITY VALUES

The value 980.155 gals published by the U. S. Coast and Geodetic Survey for station ✱589 is tied to the Potsdam system of relative gravity values through the primary base for this country in the basement of the Commerce Building in Washington, D. C. The error of observation of ✱589 is ±1 milligal so that the absolute value of any station in this survey is no more precise than this. However, as far as geological interpretation is concerned gravity differences are important. These are considerably more precise than the absolute gravity values. Estimates of the accuracy of the field observations give a measure of the accuracy of the gravity differences. An analysis of 38 reoccupations of base stations shows the average variation to be .78 scale divisions, or approximately 0.3 milligals.

LOCATIONS AND ELEVATIONS OF GRAVITY STATIONS

The remaining primary data of the survey are the latitudes and longitudes and the elevations of the stations.

This area is adequately covered by U. S. Geological Survey quadrangle maps. From these locations in latitude and longitude can be scaled to an accuracy of better than a tenth of a minute. The following maps were used in this survey:

1. Allentown
2. Allentown West
3. Easton
4. Quakertown
5. Doylestown
6. Boyertown
7. Hamburg
8. Wind Gap
9. Mauch Chunk

These maps are all standard 15-minute quadrangles, and the first three are based on surveys since 1929. They are generally considered especially good for location and elevation data. The others are sufficiently accurate for the location of such landmarks as road intersections but do not provide elevation data precise enough for this work.

Field stations have been located to minimize terrain effects, the location being made on a road intersection wherever possible since such places are commonly surveyors' turning points and hence more reliable both as to location and elevation.

Precise elevation data have been obtained for gravity stations from other surveys. For 44 stations located on State highways, the elevation of the center line of the road surface nearest the station was obtained from the Pennsylvania State Highway Department District Office at Allentown.

In the local study near Camels Hump and Pinetop 81 gravity stations were run in with a plane table and alidade. Of the remaining stations all those located on quadrangles surveyed prior to 1929 and many stations on the newer quadrangles were measured by three Paulin altimeters of 3000-foot range. It is believed that these elevations are correct to better than 5 feet. This survey was conducted as follows: Two of the instruments were operated as field instruments, and one was kept at the base to be read at 5-minute intervals during field observations. Bases were established at bench marks or other places of known elevation, and field measurements were made at points no more than 3 miles from the base. Different bases and separate surveys were tied in at every opportunity. The ties agreed to within 5 feet in all cases. The computation of sea-level elevations from the altimeter readings was similar to the treatment of the gravimeter data. That is, the altimeters were operated as relative instruments, and drift corrections included actual changes in atmospheric pressure as well as instrumental changes.

The stations whose elevations were not obtained in the above ways were read from the quadrangle maps. They are nearly all on road intersections, and all on recently surveyed areas. With the exception of the few not on road intersections the elevations are believed correct to within 5 feet.

COMPUTATION OF THE BOUGUER ANOMALY

General considerations.—The variation of gravity over a region is due to a number of causes, and to study the effect of one of them it is necessary to make proper correction for the others and compute the residuals (anomalies) due to the desired effect. This is discussed in detail by Nettleton (1940, p. 51–62), Heiland (1940, p. 135–143), and others. In geological studies it is desired to observe the effect on gravity of

horizontal and vertical variations in density of masses of rock in the earth's crust. To do this gravity is computed as precisely as required for the particular locations where it has been measured on the assumption of uniform density of the crust everywhere near the stations. The difference between this computed value and the observed gravity is the residual or "anomaly" due to density variation.

The principal variations of gravity over the earth as a whole are due to latitude and to elevation. The effect of latitude has been carefully studied, and various formulae have been computed expressing gravity at sea level as a function of latitude. (*See* Daly, 1940, p. 24–34.) In any local study such as this any of the modern gravity formulae would give equally useful sea-level gravity values. In this study the 1933 International Gravity formula was used to conform with the U. S. Coast and Geodetic Survey calculations for the region. Computed gravity at a particular place differs from that given by the formula by an amount dependent on the elevation of the station and the terrain around it. The simple elevation effect, due to distance from the center of gravity of the earth, is a decrease in gravity with increasing elevation. Correction for this can ordinarily be made by multiplying the elevation by the appropriate factor to give the proper correction in gals to the sea-level gravity given by the formula. The gravity thus computed (Free-Air gravity) is less than the actual gravity by an amount equal to the attraction of the material between sea level and the station (the Bouguer effect). The Bouguer correction is most conveniently computed by considering the terrain about the station to be a flat plain at the elevation of the station, to a first approximation. The correction for such a layer of material is approximately $2\pi\sigma hG$, where σ is the density, h the thickness (equal to the elevation of the station), and G the universal gravitational constant. Wherever the actual terrain departs from a flat plain suitable correction must be made (the terrain correction). The value of gravity obtained after applying the Bouguer and terrain corrections to Free-Air gravity is known as Bouguer computed gravity, and the difference between this value and the measured gravity is the Bouguer anomaly.

In the above-described calculations it has been assumed that the earth was free from horizontal density variation everywhere "close" to the gravity station. Actually this is never exactly the case, and the Bouguer anomaly is indicative of departure from uniform horizontal density distribution. Where an anomaly of one sign persists over a region tens to hundreds of miles wide the mass anomaly causing it is of comparable dimensions and apt to be deep-seated. Such anomalies may be reduced by applying corrections based on the principle of isostasy. There are various theories reviewed by Daly (1940, p. 36–63) of the modes of isostatic compensation that have been used extensively in gravity reductions. They all have as their basis the idea that the topography of the earth's surface tends to be maintained not alone by the strength of the crust but also by some sort of density compensation, so that higher elements of topography are composed of less dense material (the Pratt hypothesis) or are supported by a "root" of light crustal material immersed in denser subcrustal layers (the Airy hypothesis). Isostatic corrections based on either of these hypotheses or a combination of the two have been successful in reducing anomalies of regional extent all over the earth. However, in a local study of this

character the interest is not so much in broad, regional trends as in smaller local features due to surface or comparatively shallow structures. Also, each of the various isostatic corrections is admittedly arbitrary, and if used in local studies may contain computational inaccuracies that might accumulate to reverse otherwise valid local trends. For this reason the Bouguer anomaly is to be used in a local study, and if the local picture is apparently superposed on a regional trend its character can be ascertained and corrected for in a fashion similar to the correction for the regional thickening of beds in geologic structural mapping.

Bouguer anomaly computations.—In calculating the "computed Bouguer gravity" theoretical gravity at the proper latitude was read from the 1933 international gravity formula. The elevation and Bouguer corrections are both proportional to elevation and hence may be combined. The resulting correction is always negative and equal to the elevation of the station in feet above mean sea level multiplied by the factor 0.05960×10^{-3} gals./ft., the proper factor for an assumed density of 2.7 gm./cm.^{3} For most of the stations a small negative terrain correction had to be applied. Thus the Bouguer computed gravity is obtained by subtracting the sum of the terrain correction and the combined elevation and Bouguer correction from theoretical gravity. Finally, computed gravity is subtracted from observed gravity to obtain the Bouguer anomaly, which is negative throughout this region. Table 7 lists all stations occupied, their latitude, longitude, and elevation, and the Bouguer anomaly computations.

Assumed density for the Bouguer correction.—The density of country rock assumed in the Bouguer correction should be the average density of surface rocks near the station. In this region densities range from 2.3 to 3.0 gm./cm^3., but mainly between 2.6 and 2.8 gm./cm^3. These are principally the Byram granite, the Paleozoic limestones and shales, and some of the Triassic sediments. Accordingly a density of 2.7 gm./cm^3. was assumed in the Bouguer reductions, and the resulting anomaly variations are due to (1) departures from the assumed density down to sea level, and to (2) departure from uniform horizontal density distribution below sea level.

Terrain corrections.—In this region terrain corrections are generally small, and for many stations in the limestone valley they are less than 0.1 milligal. Terrain corrections were made from Hammer's (1939, p. 184–194) table and are carried out to zone M.

Inasmuch as the uncertainty in observed gravity is 0.3 milligal and the elevations of many of the stations may be 5 feet off it is useless to make terrain corrections accurate to better than 0.1 or 0.2 milligal. Therefore the following procedure was adopted. The region was divided into areas of similar terrain. These were:

(1) Triassic lowlands
(2) Highlands of the Reading Prong
(3) Flat valleys of the Reading Prong
(4) Limestone valley
(5) Sandstone and shale hills
(6) Region near and on Kittatinny mountain

Many stations were reduced in detail in each area. Except in the highlands of the

Reading Prong and near Kittatinny Mountain the stations in one area differed from each other by no more than 0.2 milligal so long as the contribution through zone E (outer radius 1280 feet) was not over 0.1 milligal. The stations reduced in detail were so picked that they covered all the areas not so much with even spacing as with representative spacing. For example, when it was found that terrain corrections on

TABLE 4.—*Terrain corrections for stations located in areas of similar terrain*

Area	Terrain corrections (in milligals)	
	Lowest	Highest
1. Triassic Lowlands	0.0	0.5
2. and 3. Reading Prong	0.2	1.2
4. Limestone valley	0.0	0.2
5. Sandstone and shale hills	0.2	0.4
6. Region near and on Kittatinny Mountain	0.5	3.0

many stations in the central portion of the limestone valley were less than 0.05 milligal closer attention was paid to stations near the slope of South Mountain to the south and the shale belt to the north. Similar procedure was followed in the other flat-lying areas. From the results a rough pattern was evident from which estimates of the terrain corrections for all other stations in these areas were made. Many of these estimates were checked by detailed reductions, and none was found to be in error by as much as 0.1 milligal.

All stations within 2 miles of the crest of Kittatinny Mountain were reduced in detail. In a few, particularly ✱678 at Palmerton, steep slopes are so close that it is questionable whether the correction is very accurate. In general in this area, however, corrections are believed to be accurate to better than 0. 2 milligal.

In the Reading Prong area several methods were used. On the gentler slopes it was justifiable to estimate corrections after a number of detailed reductions had been made. Likewise, in the limestone valleys, such as Saucon valley, estimates could be made. However, on South Mountain itself practically all stations whether on top of the hills or down on the slopes were reduced in detail. The only exceptions to this rule were a few stations the corrections for whose outer zones could be interpolated from near-by stations at nearly the same elevation.

In spite of giving terrain corrections minimum consideration their uncertainty is small compared with probable elevation errors, for example, and in general are not off by more than 0.2 milligal. Table 4 reviews the range of the terrain correction for the various zones of the survey and shows that failure to make the terrain correction would not have changed the general character of the anomalies except in the Reading Prong and on Kittatinny Mountain.

RESULTS

The Bouguer anomaly has been computed for all 712 stations established except 14 stations near Leithsville and Springtown. These 14 stations were not located on road intersections, and their elevations have not been measured inasmuch as the

area was adequately covered for the purposes of this survey without their use. Gravity values and descriptions of the locations are available for these stations for use in any more detailed work.

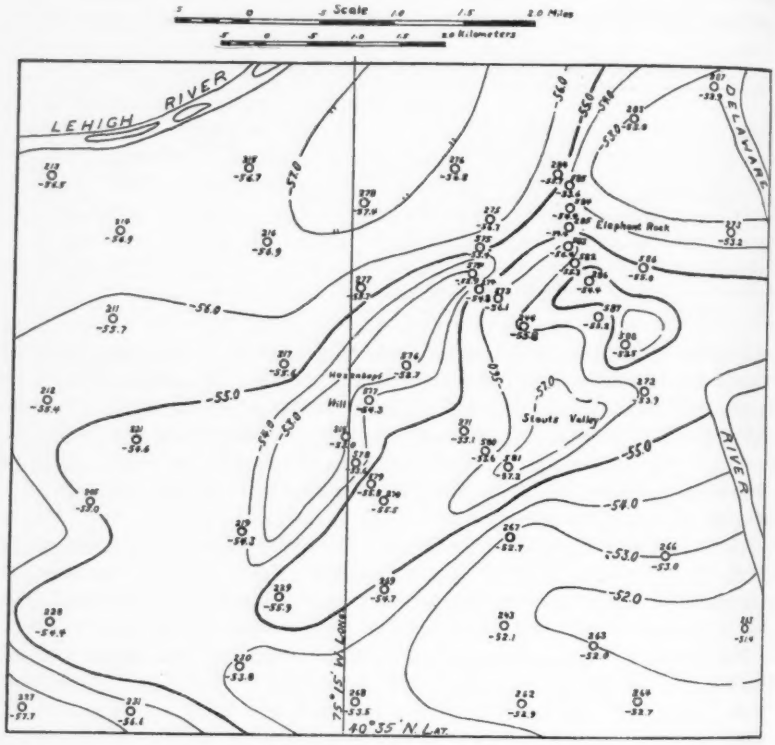


FIGURE 3.—*Hexenkopf gravity map*

Contour interval, 1 milligal. Gravity stations are indicated by circles with the number of the station north and the Bouguer anomaly (in milligals) south.

The locations and Bouguer anomalies of most of the stations are plotted on Plate 1. Variation of the Bouguer anomaly over the region is shown by means of contour lines with a contour interval of 1 milligal. Over much of the region the stations are in a grid arrangement, as regular as the road system would permit. In the outer areas the control is by a series of roughly parallel profiles. In the central and eastern portions of the region are some particularly interesting local anomalies. In these localities some stations were established 0.1 mile apart to prove the existence of the anomalies and define their shape and extent more precisely. Large-scale Bouguer anomaly maps were drawn for these localities (Fig. 3; Pls. 2, 3). On Plate 3 the Bouguer anomalies, accurate to within 0.2 milligal, are more precise at least in their

relation to one another than any others in the survey. On Plate 2 and Figure 3 there are a few stations without sufficiently accurate elevation data. Eight such stations occur on Plate 2, and six on Figure 3. Nevertheless, the general relations shown on these maps are not impaired by the uncertainty in the elevations of these stations.

ACCURACY OF RESULTS

The uncertainty in observed gravity, elevation data, and terrain corrections is responsible for the largest errors in the data and computations. The uncertainty in observed gravity differences is ±0.3 milligal. With an uncertainty of 5 feet in much of the elevation data the uncertainty in the Bouguer anomaly from this source is about 0.3 milligal. Terrain corrections are generally accurate to ±0.2 milligal except on Blue Mountain where the errors of reading in the zones near the station are likely to introduce a larger error. Therefore, individual relative anomalies may be 0.8 milligal off, and a few may, of course, have larger errors. However, the average uncertainty is about ±0.4 milligal.

The Bouguer anomaly computed by the U. S. Coast and Geodetic Survey contains additional corrections which are not included in these computations since they are small. In calculating Bouguer computed gravity the Bouguer correction used treated the terrain about the station as a flat plain to a first approximation. This was corrected for the actual terrain to a distance of 71,996 feet (zone M). Beyond this corrections arise, due to the curvature of the earth and further terrain effects. These corrections are about 0.7 milligal for the stations on Kittatinny Mountain and at most 0.2 to 0.3 milligal for stations under 1000 feet elevation in this region. Even though these corrections are included by the U. S. Coast and Geodetic Survey and not included in this work, near-by anomalies from the two surveys are comparable. Some of the U. S. Coast and Geodetic Survey stations are plotted on Plates 1, and 3, and Figure 3 for comparison.

GEOLOGIC INTERPRETATION OF GRAVITY ANOMALIES IN CENTRAL-EASTERN PENNSYLVANIA

GENERAL STATEMENT

Plates 1, 2, and 3, and Figure 3 show the variation of the Bouguer anomaly over the survey. The degree of correlation between the gravity evidence and regional and local geology is discussed in the following section. To provide a background for this discussion the geology of the region is reviewed, and a brief discussion of the methods used in analyzing the geophysical results is presented.

GEOLOGY OF CENTRAL-EASTERN PENNSYLVANIA

Stratigraphy.—Within this area there are pre-Cambrian, Lower Paleozoic, and Triassic rocks. In places these are covered by a thin veneer of glacial debris, which is nowhere thick enough to figure in this gravity survey. Table 5 gives the stratigraphy of the area.

Physiography.—The area is divided into three provinces: the Triassic Lowlands to the south, the Reading Prong of the New England highlands, and the Valley and Ridge province of the folded Appalachians on the north. The region is in a late mature stage, so that the hills are well rounded, and, wherever the underlying rock

is limestone or shale, the topography is flat or gently rolling. This permits easy location of gravity stations where the terrain correction is small.

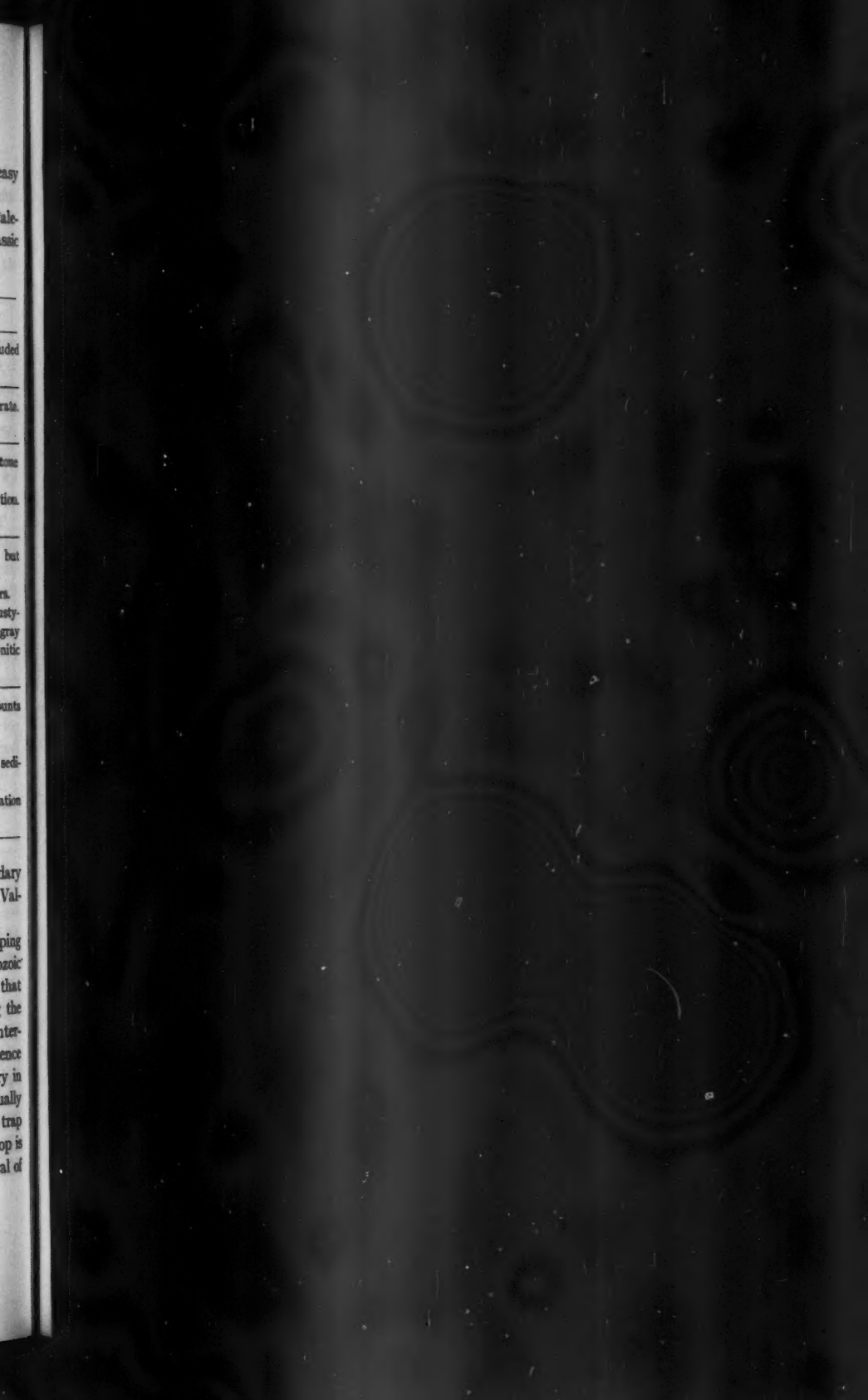
Geologic structure.—Complexly faulted and folded pre-Cambrian and Lower Paleozoic rocks of the Reading Prong separate the northeastern boundary of the Triassic

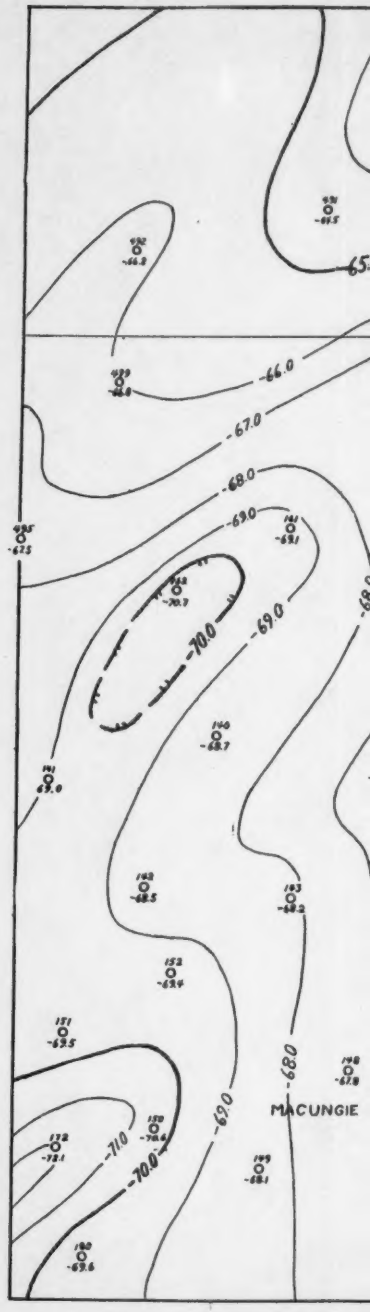
TABLE 5.—*Stratigraphy of central-eastern Pennsylvania*

Series	Formation	Lithology
Triassic	Newark	Red beds of shale and conglomerate, and intruded diabase.
Silurian	Tuscarora (Shawangunk)	Fine- to coarse-grained sandstone and conglomerate.
Ordovician	Martinsburg	Gray to black fissile shale and slate, with sandstone members.
	Jacksonburg	Argillaceous limestone of highly variable composition.
	Beekmantown	Dense, dark limestone, highly dolomitic.
Cambrian	Allentown	Dense, gray limestone, usually highly dolomitic, but of variable composition.
	Tomstown	Highly dolomitic limestone with local shale members.
	Hardyston	A sedimentary quartzite, limonitic, showing a rusty-brown weathered surface, but commonly light gray on a fresh surface. Locally grades into limonitic shale.
Pre-Cambrian	Byram	Granite or granite gneiss, containing variable amounts of hornblende and other mafic minerals.
	Pochuck	Basic gneiss of probable igneous origin.
	Moravian Heights	Quartz-sericite-sillimanite schist of probable sedimentary origin.
	Franklin	Generally a graphitic limestone, with local silicification and silication.

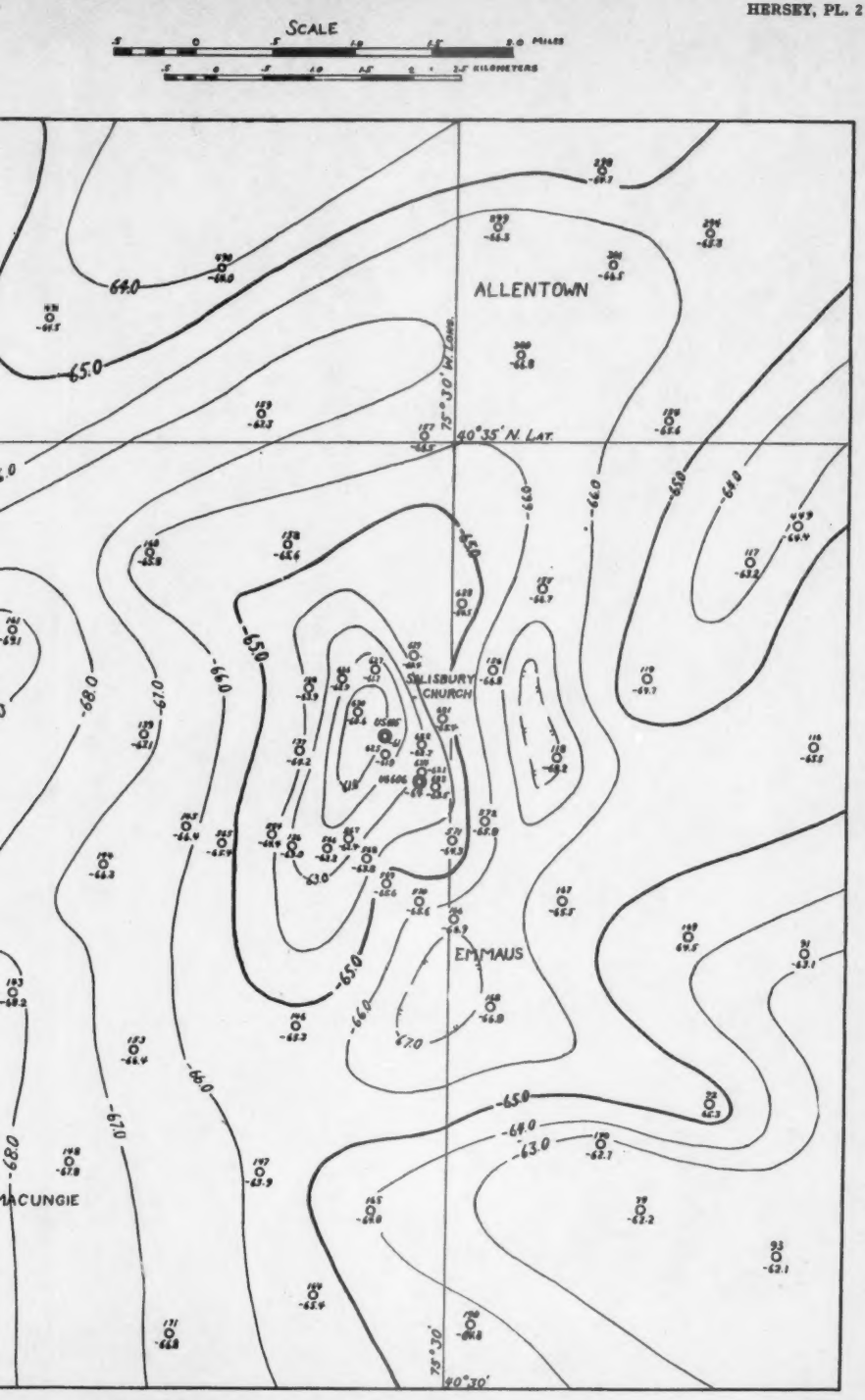
from the limestones and shales of the Appalachian valley. The northern boundary of the area is Kittatinny Mountain, which forms the first prominent ridge of the Valley and Ridge province of the folded Appalachians.

The Triassic exposed in this area is largely Brunswick shale, generally dipping gently to the north or northwest. The contact with pre-Cambrian and Paleozoic rocks to the north is not clearly exposed in this area. Some geologists believe that this contact is a major normal fault system such as is clearly observed along the northwestern boundary of the Triassic at various places in New Jersey. This interpretation is shown on the Pennsylvania geologic map. Others doubt the existence of major faulting right at the boundary but place it southeast of the boundary in the Triassic. The Triassic is intruded by large bodies of basic igneous rock, usually diabasic and commonly known as trap rock. The northernmost exposures of trap rock lie within 0.5 to 2 miles of the Triassic-Paleozoic contact, and their outcrop is horseshoe-shaped opening northwestward toward the contact. There are several of



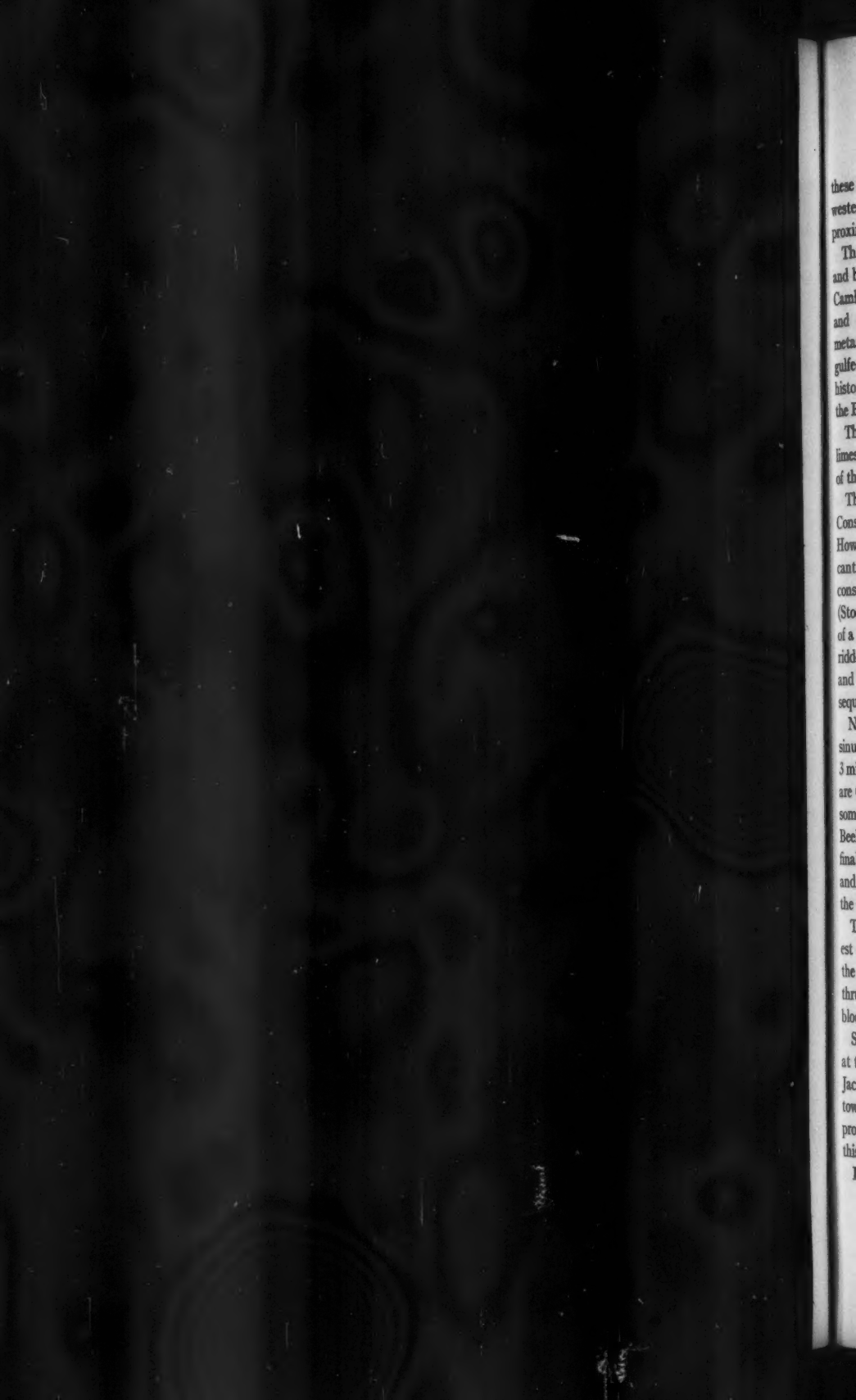


Contour interval, 1 milligal. Gravity stat
(in milligals) south.



SALISBURY CHURCH GRAVITY MAP

Gravity stations are indicated by circles with the number of the station north and the Bouguer anomaly



these structures side by side in a northeast-southwesterly trend paralleling the northwestern Triassic boundary. In this survey only the one which has its center approximately at Quakertown has been studied.

The Reading Prong is comprised of low hills, the highest being just over 1000 feet, and broad valleys from 300 to 400 feet above sea level. The major hills are of pre-Cambrian rock—Byram granite gneiss, Pochuck gneiss, and the Moravian Heights and Franklin formations. The Pochuck, Moravian Heights, and Franklin are metamorphosed older igneous and sedimentary rocks, intruded and partially engulfed by a granitic magma which crystallized to form the Byram. The complete history of this intrusion is very complex, and there is evidence of later intrusion into the Byram, though this may have occurred very soon after the Byram was emplaced.

The broad valleys are generally underlain by Cambrian and Lower Ordovician limestones, with Hardyston quartzite cropping out around the borders of many of them.

The structures both in the limestone and the pre-Cambrian are extremely complex. Considerable folding and high-angle faulting parallels the Appalachian trend lines. However, the complexity of the structures further confused by the paucity of significant outcrops so obscures the relations between the rock units that there is still considerable controversy even as to the major structural relations. In one view (Stose and Jonas, 1939, p. 757–780) the pre-Cambrian hills are the erosional remnant of a huge overthrust sheet, and the limestone valleys are windows exposing the overridden limestone. According to others (Miller and Fraser, 1939, p. 293–303) folding and high-angle faulting is dominant, and the formations are exposed in normal sequence so that the limestone valleys are major down-folds into the pre-Cambrian.

North of the Reading Prong lies a strip of Cambrian limestone following the sinuous northern front of South Mountain. North of Bethlehem this strip is about 3 miles wide and terminates against a line of low ridges, the most prominent of which are Camels Hump and Pinetop. These hills are largely pre-Cambrian rock but have some Hardyston quartzite exposed on their flanks. To the north is a broad area of Beekmantown limestone overlain by the Jacksonburg (cement rock) limestone and finally the Martinsburg shale and slate. The lower Silurian Tuscarora sandstone and Shawangunk conglomerate are the youngest Paleozoic rocks in the area, forming the crest of the ridge of Kittatinny mountain.

The Cambrian limestone strip and the Camels Hump trend are of particular interest structurally in their relation to the pre-Cambrian to the south. According to the overthrust theory Camels Hump would be another cut-off remnant of the overthrust sheet. According to the folding-high-angle fault theory it would be a fault block of older rock bounded by rather steep fault zones.

Some interesting minor structures occur in the Beekmantown area, particularly at the National Cement Company property north of Camels Hump, where areas of Jacksonburg cement rock are apparently downfaulted into the dolomitic Beekmantown. There are also several occurrences of islands of older limestones apparently protruding through younger formations over areas 1 or 2 miles across elsewhere in this part of the Appalachian Valley.

In the Martinsburg area the stratigraphic relation of the Shochary sandstone

member of the Martinsburg to the shale and slate presents an interesting problem. This sandstone forms prominent ridges in Lehigh County and extends in a rather narrow strip northeastward toward the Delaware River within the area of the survey. To the west of the area studied these ridges clearly occupy the centers of typical long, narrow Appalachian-type synclines. Willard (*in* Miller *et al.*, 1941, p. 213–228) has studied the relation of the sandstones to the shale and slate in Lehigh County and has concluded, on stratigraphic evidence, that the same situation obtains there, so that the sandstone member in that area is younger than the shale and slate. This area was included in the gravity survey to determine what, if anything, was to be concluded from the gravity picture regarding the structure in the Martinsburg.

Density measurements.—In any gravimetric survey the density of rocks involved must be so determined as to represent the effective density of a large volume of the rock in place. Here the densities were determined by weighing hand specimens in air and in water. In order to obtain proper effective densities several samples from a single locality were checked against each other and with similar suites from the same formation at other places in the survey.

In mature topography fresh specimens are hard to obtain unless there are recent road and railroad cuts and quarries. These sources have been heavily relied upon, and many other possible sources of fresh samples have been visited. Even so it is felt that some of the suites of samples are unreliable or, at best, questionable because they are somewhat weathered. However, it is believed that these measurements as a whole have indicated representative densities that are useful for interpretation. The results of the density measurements are discussed in the following.

(1) The Pochuck gneiss in its "pure" phase has an average density of 3.0 gm./cm^3. In samples that show evidence of invasion by the Byram the density varies from that figure down to 2.65 gm./cm^3.

(2) The Byram gneiss has an average density of 2.65 gm./cm^3., the variation in individual samples being about 0.02 gm./cm^3. A few samples of Byram containing little or no mafic minerals measured 2.61 gm./cm^3. These were obtained at the western end of Pinetop.

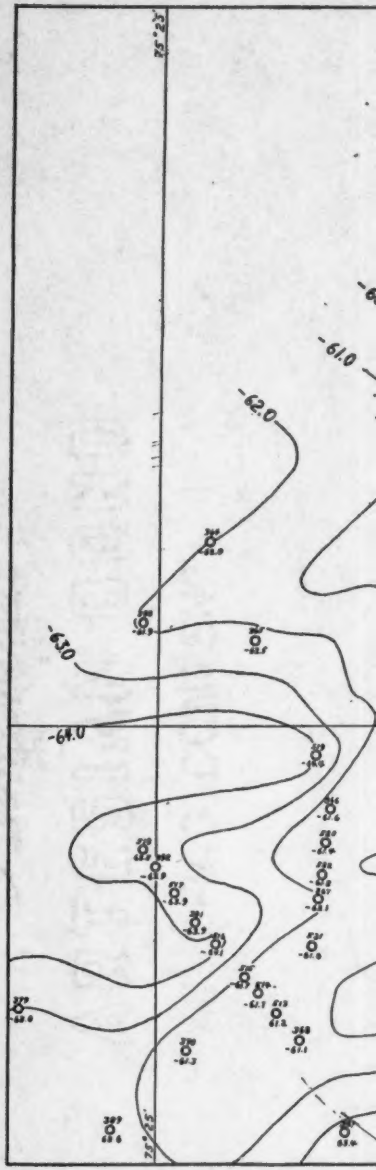
(3) The Hardyston quartzite is somewhat variable, an average of 2.63 gm./cm^3. being representative.

(4) The Tomstown and Allentown limestones are variable, ranging from 2.63 to 2.80 gm./cm^3.

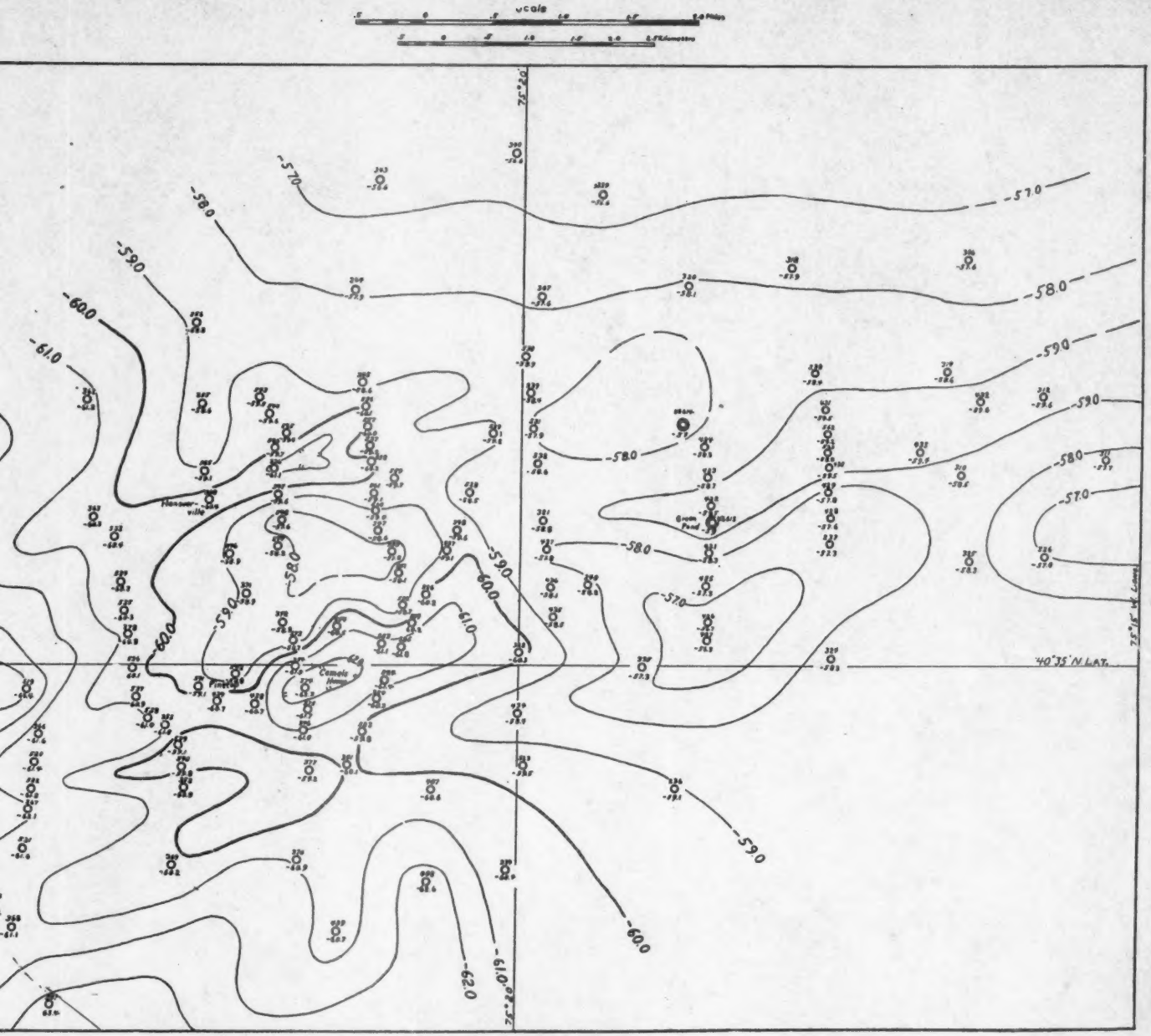
(5) The Beekmantown dolomite measures consistently close to 2.8 gm./cm^3.

(6) The Jacksonburg limestone is highly variable, ranging from 2.54 to 2.71 gm./cm^3. The bulk of samples are grouped between 2.65 and 2.70, so that 2.68 gm./cm^3. is a useful average.

(7) The Martinsburg shale samples are extremely variable, but much of the variation is due to weathering. Slate samples have densities close to 2.77 gm./cm^3., and fresh samples of shale have values quite as high in many cases. The Shochary sandstone member of the Martinsburg is likewise variable, many of the weathered samples measuring between 2.0 and 2.4 gm./cm^3. Nevertheless, unweathered samples of density 2.60 to 2.65 gm./cm^3. are common, so that an average value of 2.64 is reasonable. Some samples of dense, dark-gray pyritic sandstone from a road cut south of Slatington have a density of 2.72 gm./cm^3.



Contour interval, 1 milligal. Gravity statio



CAMELS HUMP-PINETOP GRAVITY MAP

avity stations are indicated by circles with the number of the station north and the Bouguer anomaly (in milligals) south.

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(8) In the Triassic lowlands the trap rock has a density uniformly close to 3.0 gm./cm³. The sediments, largely Brunswick shale and the fanglomerates, are variable. The density of the Brunswick shale varies from 2.33 to 2.8 gm./cm³. It is likely that most of the values below 2.55 are due to weathering, since few localities have absolutely fresh red shale. Samples from the high cliffs along the Delaware River southeast of Durham are uniformly about 2.8, and elsewhere in the deep railroad cuts and other probable fresh outcrops densities range from 2.62 to 2.71 gm./cm³. A value of 2.65 gm./cm³. is assumed to represent the average for all Triassic sediments in the area.

INTERPRETATION OF GRAVITY ANOMALIES

General considerations and methods.—The variation of the Bouguer anomaly over a region may be regarded as made up of a regional variation with superimposed local variations. The regional variation, or anomaly, may take the form of a uniform decrease in the anomaly in a given direction, or it may follow a somewhat more complex but regular pattern over the region. The regional anomaly is due to some large deep-seated mass anomaly the influence of which is felt over a wide area at the surface, or to a shallower mass anomaly that persists throughout the region. The first might be due to failure to make the isostatic correction near a major mountain system or near a continental margin, while the second might be the increase in thickness of light sediments over denser basement rocks going into a basin area. Superimposed on these more gradual changes are effects of mass anomalies of relatively lesser horizontal extent which are shallow enough to have measurable effect on gravity at the surface.

Where the interest is principally in local effects the regional anomaly is to be removed. This is done in a fashion similar to the correction for regional thickening of sedimentary rocks in geologic structural mapping. The regional correction to be applied is obtained by some suitable averaging method applied to the data over the surrounding region.

Local anomalies are to be correlated with the known geology of the region. This stage in the analysis includes comparison of general structural trends with the trends of the anomalies, geographical correlation of known or suspected mass anomalies with the gravity features, and finally rough or detailed analysis of the gravity anomalies to determine what density contrasts and shapes and depths of mass anomalies might cause them. The detail justified in this last stage of the analysis depends not only on the quality of the gravity data, but also on the control other than gravity available for the problem. Gravity effects are proportional to the density contrast and size of the mass anomalies involved, and inversely proportional to the square of the distance to the center of gravity of each anomalous element. If none of these factors is known from other evidence, there is little means of delimiting the shape, size, or depth of the mass anomaly causing the gravity anomaly. Nevertheless, the gravity data plus some rather simple considerations limit the depth and size to ranges that are useful for preliminary study. Obviously a gravity anomaly must be larger in horizontal extent than the mass anomaly producing it. This gives an upper limit to the possible size of the mass anomaly, though it may be any size smaller than this limit as far as this consideration is concerned.

Anomalies due to simple forms.—Comparison of the actual anomaly with anomalies computed for simple geometrical forms helps to fix the limiting depth of the center of gravity of the anomalous mass. This has been discussed by Nettleton (1940, p. 122–125) who remarked that "if little or no other control is available, the comparison of the effects of such simple forms with those observed may give as close a

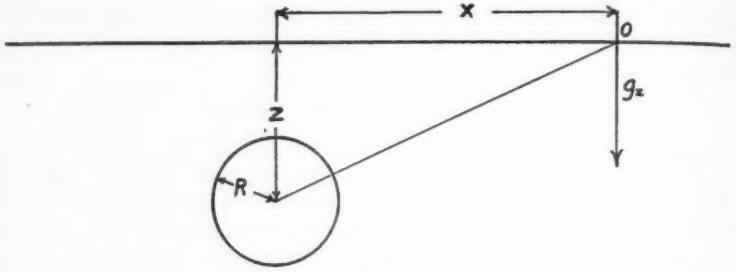

FIGURE 4.—*Section of a spherical or cylindrical mass anomaly*

fit to the observed data as justified by other uncertainties of the problem." The portion of his analysis pertinent to this survey is outlined below for the sphere, the horizontal cylinder, and the semi-infinite horizontal slab, or fault.

THE SPHERE: Vertical gravity due to a sphere of radius R, having a density contrast σ with its surroundings is given by

$$g_z = \frac{4\pi G\sigma R^3}{3} \frac{z}{(z^2 + x^2)^{3/2}}$$

where z is the depth to the center of the sphere and x is the horizontal distance along the surface from a point directly over it (Fig. 4), and G is the universal constant gravitation.

THE HORIZONTAL CYLINDER: Vertical gravity due to an infinitely long horizontal cylinder is given by

$$g_z = 2\pi G\sigma R^2 \frac{z}{z^2 + x^2}$$

where the symbols have the same meaning as in the case of the sphere with the exception that x is the horizontal distance from a line on the surface directly above the cylinder, and R is the radius of its cross section.

THE SEMI-INFINITE HORIZONTAL SLAB, OR FAULT: The mass anomaly is a slab of material of thickness t, density contrast σ, whose average depth is z, and terminates along a vertical plane (Fig. 5). Vertical gravity at o due to this slab is

$$g_z = 2G\sigma t \left(\frac{\pi}{2} + \tan^{-1}\frac{x}{z}\right)$$

Depth estimation.—The gravity anomaly across these mass anomalies is shown graphically in Figures 6 and 7. From these it is seen that the half width of the

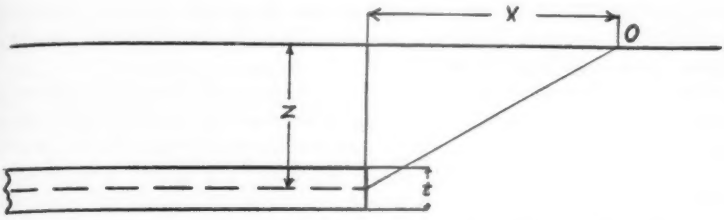

FIGURE 5.—*Mass anomaly due to vertical faulting—a semi-infinite horizontal slab*

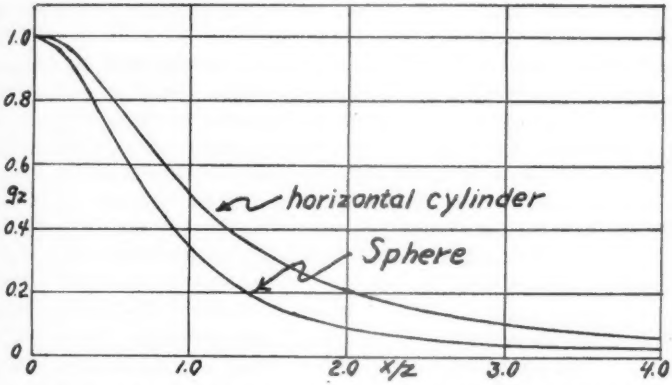

FIGURE 6.—*Gravity anomalies over a sphere and an infinite horizontal cylinder*

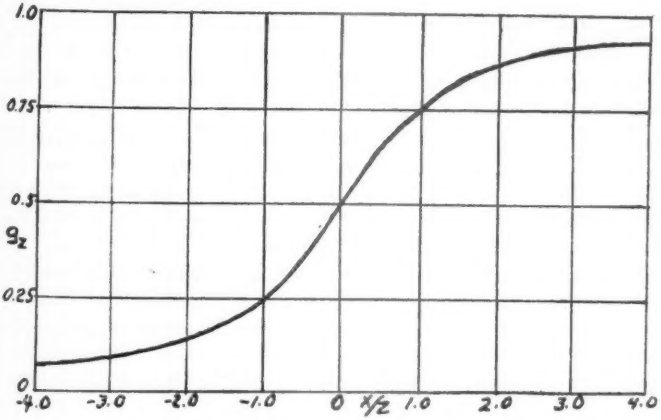

FIGURE 7.—*Gravity anomaly over a vertical fault*

anomaly at half its maximum value is 0.766 z for the sphere, z for the cylinder, while for the fault the anomaly is half its value from the fault trace to infinite distance at a distance $x=z$ away from the fault trace.

Where anomalies are found that show a roughly circular shape the width at half-maximum may be measured. The depth of the center of gravity of an equivalent sphere is 0.653 of this width. Similar comparisons can be made for long relatively narrow anomalies, and the depth of the equivalent cylinder or semi-infinite slab computed. Such computations are only approximate, and, further, they give maximum depths since wider and shallower mass anomalies could produce the same gravity anomaly.

For more detailed analysis data other than gravity is required. If, for example, the probable density is suspected the size of the equivalent simple form and the depth to its top can be computed.

If still other pertinent data are available more detailed analysis may be justified. For example, in this survey the density contrasts crop out in some cases, thus defining the depth and shape of the top of the anomalous mass to some extent, as well as leaving no doubt about the nature of the density contrast. Similar control is given when the depth of the top of buried mass anomalies is known at one or more points from well logs or seismic data. Magnetic survey data may prove useful by showing whether gravity anomalies are also magnetic. This can be used to distinguish between magnetic and nonmagnetic density contrasts. Possible use of magnetic data in this area is discussed below.

Detailed analysis of mass anomalies.—Where detailed analysis of mass anomalies is justified it is done by setting up arbitrary mass anomalies judged to approximate the true picture. The gravity anomalies produced by such a mass anomaly are computed and compared with the measured anomalies. This process is repeated until a fit with the measured anomalies is obtained that is as good as warranted by the data. How much reliance is to be placed on the best "fit" depends not only on the accuracy of the computations and the gravity data, but also on the control other than gravity that served as a background for picking the mass anomalies analyzed.

There are several graphical and mechanical computational methods developed for this type of analysis (Nettleton, 1940, p. 115–119; Heiland, 1940, p. 143–157; Breyer, 1938, p. 317–336). One of these is a scheme employing a transparent circular template on which a number of dots are laid out in regular arrangement about the center. Each dot represents a known contribution to gravity at the center of the pattern. By placing the template with its center at the gravity station and then counting the dots that fall within the section of the assumed mass drawn to proper scale the gravity anomaly is obtained simply by multiplying by an appropriate scale factor.

Such a "dot chart" has been used for rough computation of anomalies over the Triassic trap rock and Shochary Ridge. The chart was designed to give the anomaly due to infinitely long horizontal mass elements from their vertical section. The error involved in using such a chart without correction for the finite length of the mass is about 11 per cent when the ratio of the length of the mass element to its distance from the "gravity station" is 4.0. For ratios greater than 10.0 the error is less than 2 per cent (Nettleton, 1940, p. 17). In the anomalies mentioned the

gravity relief is about 7 mg. for the trap rock and a maximum of about 2 milligals at Shochary Ridge. Since the mean error of the Bouguer anomalies is 0.4 milligals and the ratio of length-to-distance for the mass elements is almost everywhere between 2 and 10, and seldom less than 4, any more refined treatment than the assumption of infinite length is not justified.

INTERPRETATION OF THE RESULTS OF THIS SURVEY

Regional anomaly.—Plate 1 shows the general features of the Bouguer anomaly variation to be considered first in relation to what is known of the gravity field over the eastern United States.

From data already published by the U. S. Coast and Geodetic Survey this area is known to lie at the northeastern end of an oval-shaped gravity "trough" trending east-northeast west-southwest and extending from Easton to Harrisburg, Pennsylvania. In this region gravity gradients are low—about 1 milligal per mile—and agree with results obtained by smoothing off the contours of Plate 1. Plate 1 shows a more highly negative area lying to the west-southwest where there are Bouguer anomaly values as low as −73 milligals. Elsewhere in the trough no anomalies lower than −68.5 milligals occur, so the area centering about Lock Ridge to the southwest of Allentown is perhaps the lowest portion of the trough. The problem of the significance of the negative trough is beyond the scope of this investigation. However, the smaller local features that are characteristic of gravity anomalies in this region show a close relation to the known geology and frequently show a definite correlation with the areal geology.

Density contrasts.—Density contrasts likely to produce anomalies are tabulated in Table 6 with the areas in which they may be expected to have significant effect.

In cases where the correlation between areal geology and gravity anomaly is definite and clear the nature of the density contrast is clear also. The best example of this is the high correlated with the Triassic trap rock. Another is the gravity high at Hexenkopf Hill which is made principally of Pochuck gneiss. In some other cases the relations are not so clear cut, and other evidence is needed to determine the nature of the density contrast. In the case of the Camels Hump-Pinetop area magnetic measurements indicate certain gravity highs to be highly magnetic. In general, this would imply that the anomalous body is a basic igneous or metamorphic rock having a higher magnetite content than the surrounding more acidic igneous rocks or sediments. The known associations of the Byram and Pochuck indicate that this rule holds, and a gravity feature that shows definite correlation with a magnetic anomaly is here taken to be due to a mass anomaly in the pre-Cambrian section. Since only one small portion of the area has been completely surveyed magnetically no data are available for use with most of the gravity anomalies whose density contrasts are not known.

Local features.—The Triassic, pre-Cambrian-Cambrian, and Ordovician areas present different gravity pictures. The Triassic is dominated by the large northeast-southwest-trending feature that centers about Quakertown. In the Reading Prong there is a regional gradient slightly south of east of about 1 milligal per mile. Superimposed on this are several long, narrow anomalies that parallel the structural trends, and in the east at Hexenkopf Hill a large positive anomaly correlates with an

extensive area of Pochuck gneiss. Smaller areas of Pochuck are near the high parts of other gravity "ridges" such as the arcuate ridge south and east of Allentown. A small area of Pochuck is mapped 1 mile east of the sharp bend in the Lehigh River at Allentown on the north side of the river, which corresponds exactly with the high

TABLE 6.—*Density contrasts*

Area	Contrasting rocks	Density (gm./cm.³)	Density contrast (gm./cm.³)
Triassic Lowlands	Trap rock (diabase) Brunswick shale	3.0 2.55–2.65	0.35–0.45
Reading Prong	Pochuck gneiss Byram gneiss	3.0 2.65	0.35
	Limestones Byram gneiss	2.65–2.8 2.65	0.0 –0.15
	Pochuck gneiss Limestones	3.0 2.65– 2.8	0.20–0.35
Appalachian Valley	Pochuck gneiss Byram gneiss	3.0 2.65	0.35
	Limestones Byram gneiss	2.63–2.8 2.61–2.65	0.0 –0.19
	Beekmantown ls. Jacksonburg ls.	2.8 2.54–2.71	0.09–0.26
	Martinsburg shale and slate Shochary ss.	2.77 2.63	0.14

part of this anomaly. A small area of Pochuck is found within the −60.0 contour of the high at station # 194, about 8 miles due south of Allentown. These correlations suggest that these gravity highs may be caused by more extensive Pochuck masses to which these outcrops belong.

In the Cambrian limestone area just north of the Reading Prong the principal features are the Salisbury Church high and the long anomalies extending from Easton to Allentown (Pls. 2, 3). These are probably due to the same sort of basement rock mass anomalies as occur in the Reading Prong.

In the Ordovician the principal feature is the broad high just north of Bath. This follows the regional structural trends but does not correlate directly with any surface rocks. Since no other geophysical or other data are available in this part of the area further analysis of this feature is unjustified. Other local features in the Ordovician are the low, broad high north of Allentown and the long lows trending eastward from Shochary Ridge and south of Slatington. The relation of these features to basement and sedimentary structures is discussed below.

Individual local anomalies.—Several of the local anomalies will be discussed in this section, correlating gravity results with the geological and magnetic data.

QUAKERTOWN (TRIASSIC) HIGH: The most prominent single feature on Plate 1 having definite correlation with known geology is the northeast-southwest-trending high in the southeast at Quakertown. This feature corresponds exactly with the easternmost large body of diabase intrusive into the Triassic of Pennsylvania. The extensive outcrops of this body bound it, thus indicating an elongated spoon shape. The maximum anomaly across it is about 7 milligals, and its shape suggests considerable thickening toward the central axis of the intrusive. Rough dot-chart calculations have been made to determine a possible cross section for the body. In these a density contrast of 0.35 gm/cm.^{3} was assumed, and the diabase body was assumed to have uniform cross section and infinite extent parallel to its long dimension. The curve marked "measured" in Figure 8 is a profile read from the contours of Plate 1 across the anomaly about 1 mile northeast of Quakertown, but differing from the actual profile by a regional correction of about 3/8 milligal per mile. The "calculated" curve was obtained from dot-chart measurements on the section shown. This section was one of five tested and seemed to give as good a fit as was warranted by the assumed conditions and the gravity data. Accordingly the gravity anomaly could be caused by an intrusive roughly 100 feet thick near its edges and about 1800 feet thick along its axis.

HEXENKOPF HIGH: This feature (Fig. 3) exhibits close correlation with the lithology and structure in this area. Hexenkopf Hill is a relatively low, well-rounded hill on which Pochuck gneiss predominatly crops out. The same body of Pochuck extends irregularly over a considerably wider area to the west. This whole area is a pronounced gravity high. Elephant Rock, east of Hexenkopf, is of Byram material. In this section of the map the correspondence of the gravity anomalies with surface outcrops apparently fails. There is a low "saddle" in gravity anomalies underneath Elephant Rock that does correlate with the Byram mass there. However, the small highs indicated by stations #286 and #287 have no counterpart in known surface lithology and should be checked by additional gravity control and a magnetic survey. Also, the larger high northeast of Elephant Rock should be measured in greater detail. These features correspond to neither the surface outcrops nor topography and may prove to be indications of basic material in the pre-Cambrian. The low in the central portion of Stouts Valley just east of Hexenkopf is readily explainable if the Pochuck is the controlling "anomalous" rock unit, since this valley is a syncline in Cambrian Hardyston and limestones.

CAMELS HUMP-PINETOP AREA: In this area (Pl. 3) are several gravity anomalies due to different mass anomalies so close together that the fields of the separate bodies greatly overlap. For this reason it is difficult to estimate the size or lateral extent of separate gravity anomalies, which, in turn, greatly decreases the value of calculations of the depth of anomalous masses. However, it is still possible to make certain definite correlations with surface geology and structural trends.

About 3 miles north of Bethlehem a series of low ridges trends slightly north of east, parallel to South Mountain. The most prominent are Camels Hump and Pinetop, which rise about 200 feet above the surrounding flat limestone valley. All the ridges consist of pre-Cambrian rocks, largely Byram granite; the Franklin formation, Hardyston quartzite, and Tomstown limestone occur locally on the flanks of Camels Hump. To the south of these ridges lies Cambrian limestone, and to the

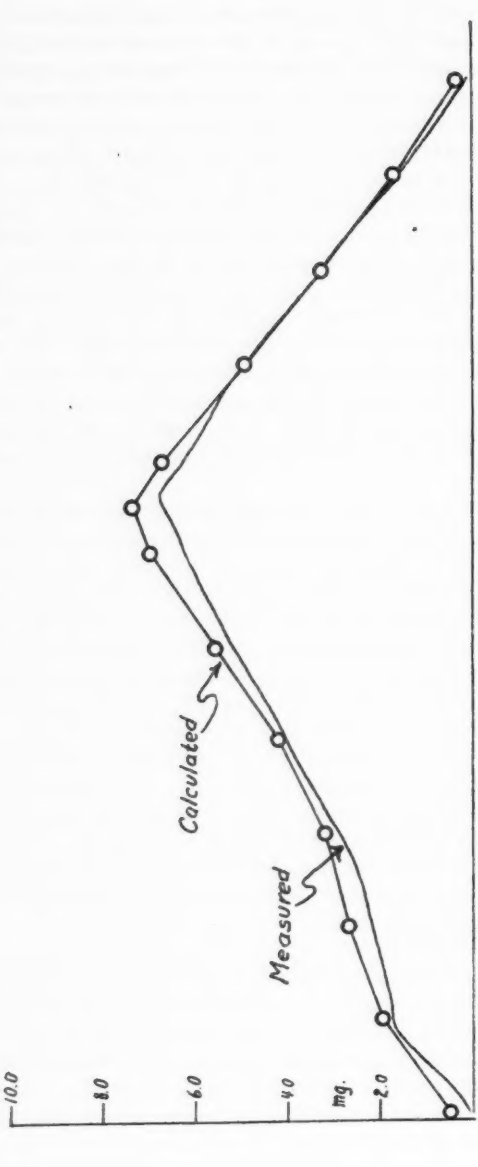

FIGURE 8.—*Gravity anomaly and section of the Triassic intrusive at Quakertown*

north, Beekmantown dolomite. The width of the outcropping pre-Cambrian nowhere exceeds half a mile across the strike. Accordingly, gravity measurements must be very closely spaced across this structure for sufficiently accurate indication of the density distribution. Many profiles using a station spacing of 0.1 or 0.2 mile were established across this area (Pl. 3).

At Camels Hump is a prominent 1.6 milligal closed low. The northern boundary of this low, taken as about −60.5 mg., follows within less than 0.1 mile the contact zone between the pre-Cambrian and the Beekmantown at Monocacy Creek. This is analyzed as a fault zone by Miller and Fraser. The gravity results indicate a denisty contrast in the position of this fault zone, reasonably assigned to a contrast between Beekmantown and Byram.

This low trend saddles across to the west over Pinetop and loses its definite character within a half a mile west of Pinetop. Continuing east-northeast on the same trend from Camels Hump a low trough centering on Green Pond carries on just northwest of Easton. This area is underlain by Beekmantown and Cambrian limestone. These is no apparent correlation of the gravity anomaly with surface lithology.

Immediately south of this low trend is a high gravity trend that extends west-southwest to the Lehigh River at Allentown and east-northeast to Easton. The persistence of this trend with the low trend to the north is in line with the structural continuity from the Camels Hump-Pinetop structure to the pre-Cambrian outcrops at Easton suggested by Miller and Fraser (1939). This gravity high trend corresponds exactly with a series of magnetic highs surveyed by Ewing and Pentz (1935, p. 186–191). This suggests the possibility that both are due to lithologic changes in the basement or to a ridge of magnetic pre-Cambrian.

One and a half miles due north of Camels Hump a downfaulted block of Jacksonburg cement rock lies completely surrounded by Beekmantown dolomite. The density contrast between the Beekmantown (2.8) and the Jacksonburg (2.65) is about 0.15. There is a gravity low of about 1.0 to 1.5 milligals that correlates well with the known Jacksonburg occurrence.

The magnetic survey of Ewing and Pentz outlined a major magnetic high of about 1000 gammas in z, the vertical component of the earth's magnetic field. The high point lies about half a mile northwest of station # 363, west of Hanoverville. This high might be produced by a south-seeking pole buried about 2.3 to 2.8 miles. A gravity high is located 1 mile north of Camels Hump and 2 miles east-southeast of the center of the Hanoverville magnetic high. Any comparison of these two surveys is greatly complicated by gravity effects due to known near-surface mass anomalies —the Byram of the Camels Hump-Pinetop trend and the Jacksonburg just discussed. Also, this lies on the flank of a much larger anomaly, the high north of Bath. As the first two are not characteristically magnetic they would not be expected to produce magnetic anomalies. The relation between the magnetic and gravity anomalies here is obscure, but because of the 2-mile offset between their high points it seems unlikely that the magnetic and gravity highs are caused by the same anomalous body, though the magnetic anomaly might be produced by a strong pole associated with the body causing the gravity anomaly.

SALISBURY CHURCH GRAVITY HIGH: Less than half a mile due west of Salisbury Church and northeast of the Lehigh Country Club is a steep slope going down to the Little Lehigh River. On this slope pre-Cambrian rock crops out. Along the east-facing slope the exposed rock shows evidence of intense crushing and shearing. A fault zone is mapped along this trend by Miller and Fraser (1941). The rock is largely Byram granite, completely surrounded by Paleozoic limestone. The nearest other outcropping pre-Cambrian rocks are in South Mountain, southeast of Emmaus, about 2 miles away (Pl. 4).

The gravity map (Pl. 2) shows that this outcrop is at the high point of a high gravity anomaly, almost completely isolated from the surrounding local picture. The high is about 6 milligals and has a total width of about 3 miles measured along an east-west line. It is about 5 miles long north and south. The width of the anomaly from $\frac{1}{2}$-maximum value on the east to $\frac{1}{2}$-maximum value on the west is about 0.8 mile. The depth to the center of gravity of a sphere producing a roughly equivalent anomaly would be 0.5 mile. Such a sphere having a radius of 0.5 mile should have a density contrast about 0.35 gm./cm.^{3} to produce so large an anomaly. This is comparable with the Byram-Pochuck contrast, and apparently this is not an unreasonable explanation considering the close correlation between the outcrop of Byram and the gravity high. A detailed magnetic survey over this area would show whether the mass anomaly is also magnetic. As yet there has been no opportunity to make such a survey.

SHOCHARY RIDGE GRAVITY LOW: The Shochary sandstone member of the Martinsburg forms a series of minor ridges a few miles south of Kittatinny Mountain in Lehigh County and westward. The most prominent of these is Shochary Ridge, type locality of the Shochary sandstone. Density contrast between the sandstone and underlying Martinsburg shale is estimated as about 0.15 gm./cm.^{3} on the basis of data made questionable because of the poor quality of density samples available. The majority of sandstone samples are deeply weathered, and the balance are not representative.

A well-defined gravity low is found to correlate with the eastern end of Shochary Ridge. This ridge clearly lies in a syncline with older Martinsburg flanking it to north and south. The same sandstone continues across Lehigh County to the Lehigh River just south of Slatington. Correlating with this is a gravity low trend superimposed on the high north of Bath. This anomaly is between 1 and 2 milligals, or about that of the Shochary Ridge low. Unfortunately, gravity control is sparse in this area, and therefore this feature is not well mapped. However, it does correlate in position with the Shochary sandstone. Further geophysical data, both gravity and magnetic, will be required to determine whether or not this feature is actually to be correlated with a density contrast between the sandstone and underlying shale.

To compare geologic data with the gravity results at Shochary Ridge a dot-chart calculation was made on the following assumptions: (1) gravity measurements are to be made at a common elevation, and (2) across a long symmetrical syncline involving shale of density 2.77 gm./cm.^{3} and younger sandstone of density 2.62 gm./cm^3. The assumed geologic section and the calculated gravity profile are

shown in Figure 9. The maximum thickness of sandstone along the axis of the syncline was assumed to be 1000 feet, in accordance with estimates of workers in this region. The maximum gravity relief is 1.46 milligals for a density contrast of 0.15 gm./cm³. The measured anomaly is between 1 and 2 milligals, so that the agreement is as close as the data warrant, and a reasonable interpretation of the gravity results agrees with the geologic interpretation.

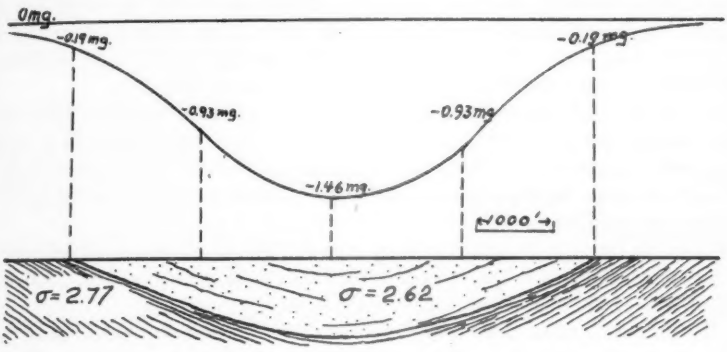

FIGURE 9.—*Gravity anomaly across a syncline*

CONCLUSIONS

The gravity anomalies have been described and correlated with the known geology and the known magnetic anomalies in this region. The conclusions to be drawn from the present study are as follows:

(1) The regional anomaly obtained by smoothing out local effects agrees with data by the U. S. Coast and Geodetic Survey, showing this area to be at the northeastern end of a major gravity low trend. The trend is known to extend from Easton to Harrisburg from their data.

(2) The shapes and trends of the smaller local anomalies parallel Appalachian structural trends.

(3) In many cases there is geographic correlation between marked gravity anomalies and surface outcroppings of density contrast sufficient to cause the anomaly (*e.g.*, the Triassic diabase at Quakertown, Pochuck at Hexenkopf, Byram-limestone contrast at Camels Hump).

(4) Two anomalies are found to have no correlated density contrast that crops out, but evidence other than gravity indicates the possible subsurface density contrast causing the anomaly. In one case the evidence now available is geological (at Salisbury Church), while in the other both geologic and magnetic evidence is available (the high trend between Easton and Allentown).

(5) The detailed study of local effects over this region generally is impeded by the somewhat complex regional variation and the superposition of several effects such as those in the Camels Hump-Pinetop area. However, some local anomalies have been usefully analyzed, and results obtained that are geologically reasonable and

fit the gravity picture (*e.g.*, the Triassic intrusive, the Salisbury Church high, and Shochary Ridge).

SUGGESTIONS FOR FUTURE WORK

In several local areas anomalies require closer control with gravity and magnetic data. Typical of these are the arcuate highs south of Allentown and Bethlehem and the more or less well-defined high running through Saucon Valley and to the southwest. The suggestion has been made that these may be due to buried Pochuck masses. Magnetic measurements over these areas should help to determine the nature of the density contrasts involved. Also, geologic data should be collected specifically for more detailed analysis of the gravity anomalies.

The gravity high northwest of Allentown should be studied in detail with a magnetic survey as well as additional gravity control. It is wide enough to be caused by a fairly deep mass anomaly and occupies a position relative to the general tectonic picture somewhat similar to the magnetic high of Ewing and Pentz, or the Salisbury Church gravity high.

Closer control and density studies of surface rocks by means of density profiles (Nettleton, 1939, p. 176–183) should be made in the Martinsburg area. Such data should help in working out structures in the sedimentary section.

The shape and trend of the large high due to the diabase near Quakertown raises two questions, both of which will require additional gravity surveying in the Triassic.

The first is the question of the shape and relation to regional structural trends of other more or less similar diabase bodies to the southwest. Do they thicken along a marked axis as this body does? If so, what pattern do these axes form in relation to other structures?

The second question is whether marked evidence of very local thickening of the diabase body can be found on this and other bodies to show possible locations of the supply ducts to the intrusive bodies. If so, what relation do these bear to the intrusive bodies? Profiles across the trend of the body with somewhat closer control are necessary.

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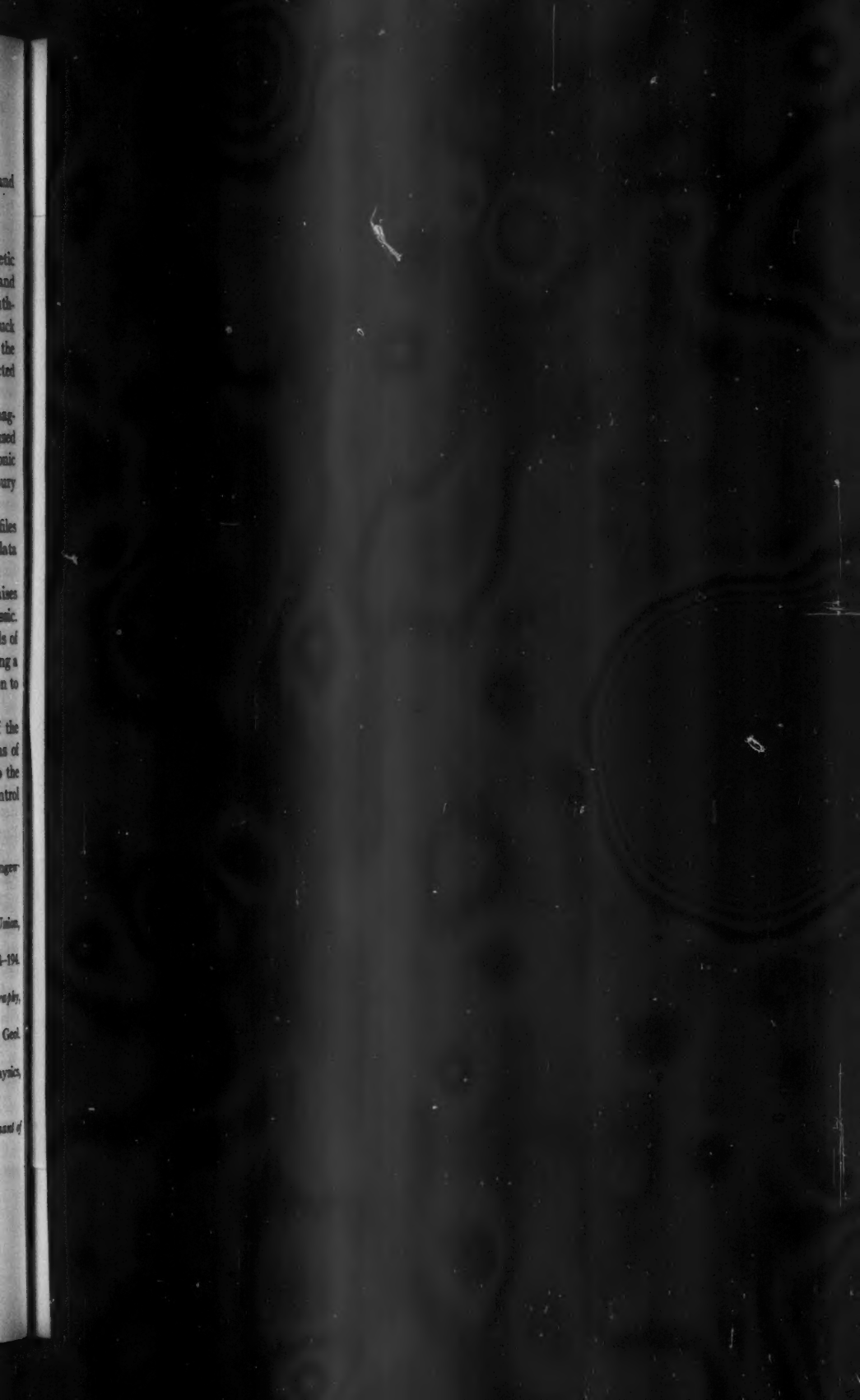
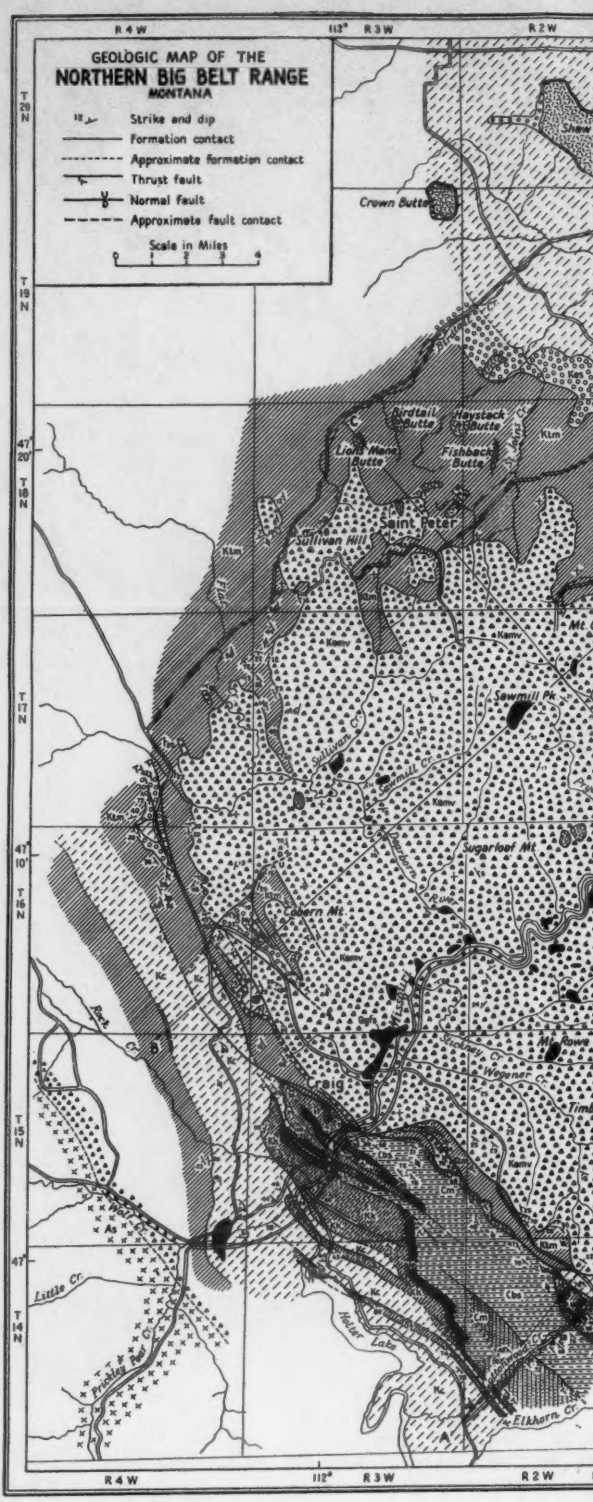


AREAL GEOLOGY OF CENTRAL-EAS

EXPLANATION
AREAL GEOLOGY

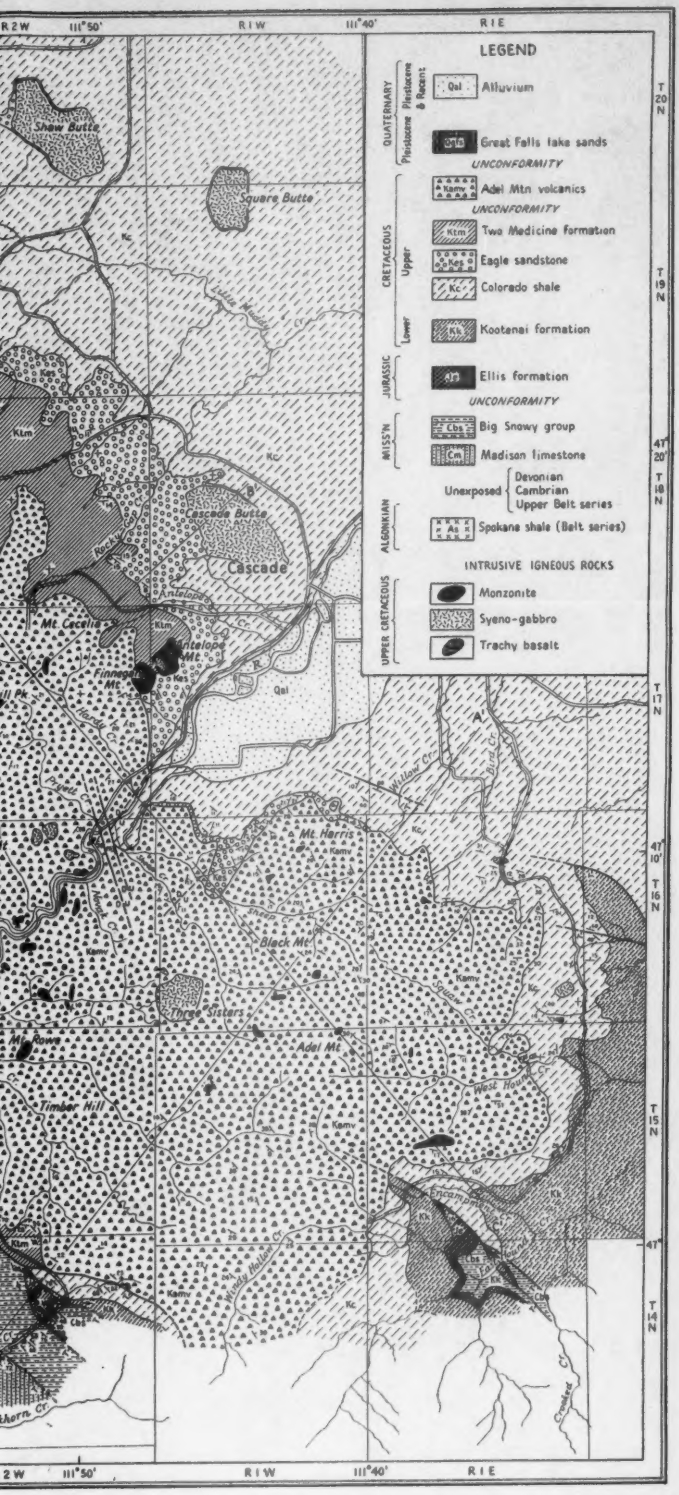
- Trd — Diabase
- Trb — Brunswick Shale — Triassic
- St/Ss — Tuscarora Cgl. Shawangunk Cgl. — Silurian
- Omu — Upper Martinsburg Sandstone and Shale
- Oml — Lower Martinsburg Shale
- Oj — Jacksonburg Limestone
- Ob — Beekmantown Limestone — Ordovician
- €a — Allentown Limestone
- €t — Tomstown Limestone
- €h — Hardyston Sandstone — Cambrian
- pC b — Byram Gneiss
- pC p — Pochuck Gneiss
- pCmh — Moravian Heights Formation
- pCf — Franklin Formation — Pre-Cambrian

RAL-EASTERN PENNSYLVANIA



STRUCTURE

GEOLOGIC MAP AND STRUCTURE SEC



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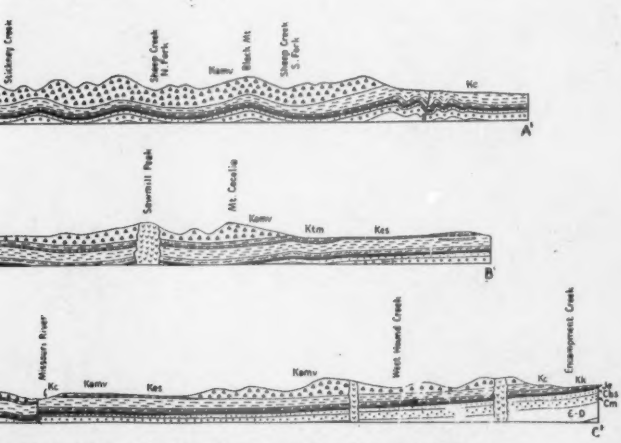

E SECTIONS OF NORTHERN BIG BELT RANGE

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IGNEOUS ROCKS OF THE NORTHERN BIG BELT RANGE, MONTANA

BY JOHN B. LYONS

CONTENTS

	Page
Abstract	446
Introduction	446
Location	446
Field work	447
Acknowledgments	447
Previous work	447
Physiography	448
Stratigraphy	449
Regional distribution	449
Spokane shale	449
Cambrian and Devonian formations	450
Madison limestone	450
Big Snowy group	450
Ellis formation	450
Kootenai formation	451
Colorado shale	451
Eagle sandstone	451
Two Medicine formation	451
Adel Mountain volcanics	452
Great Falls lake sands	452
Alluvium	452
Structural geology	452
Regional setting	452
Lewis overthrust	452
Craig imbricate zone	453
Cobern Mountain faults	453
Faults within the volcanics	454
Folds	454
Structural relations of igneous rocks	455
Introduction	455
Dikes	455
Sills	457
Laccoliths	459
Plugs and chonoliths	459
Petrography	459
Field relations	459
Plagioclase calcirhyolite	460
Quartz latite	460
Trachybasalts	460
Analcime trachybasalt and augite trachybasalt	461
Biotite pyroxenite	461

Syenogabbro 461
Hornblende monzonite 462
Augite calcimonzonite porphyry 462
Quartz monzonite porphyry 462
Monzonite 462
Sanidine monzonite 463
Biotite-augite diorite 463
Volcanic conglomerate 463
Orthoclase diabase 463
Light-colored monzonite 464
Zeolitization 464
Age of volcanics 465
Petrology 467
Chemical petrology 467
Analcime problem 469
Summary 470
References cited 471

ILLUSTRATIONS

Figure — Page
1. Map of Montana showing location of area 447
2. Columnar section of northern Big Belt Range 449
3. Map of intrusive rocks of northern Big Belt Range 456
4. Variation diagram of Big Belt, Livingston and Absaroka subprovinces 467

Plate — Facing Page
1. Geologic map and structure sections of northern Big Belt Range 445
2. Igneous structures 460
3. Photomicrographs 461

ABSTRACT

The northern end of the Big Belt Range of west-central Montana is underlain chiefly by upper Cretaceous potash-rich basalts which were extruded unconformably on Cretaceous sediments and folded with them into northwesterly trending anticlines and synclines. The western border of the igneous rocks is marked by an imbricate zone of Mississippian to upper Cretaceous rocks. The Lewis overthrust, carrying the Belt series over the upper Cretaceous strata, bounds the fault zone on the west.

The trachybasaltic breccias, agglomerates, and flows, 3200 feet thick, are intruded by many chonoliths, sills, and dikes ranging from gabbro to quartz monzonite. Some of the trachybasalts have abundant primary analcime phenocrysts. All evidence favors crystal settling as the cause for the magmatic differentiation within this petrographic province.

INTRODUCTION

LOCATION

The area described here (Fig. 1) includes about 990 square miles of the northern end of the Big Belt range and the adjacent plains. The relief in this part of the range is 3600 feet, the topography rugged, and the highest peak 7100 feet in elevation.

The region is drained by the Missouri River which flows northeasterly across the mountains, rudely bisecting them, in a steep-walled canyon. A spur of the Great Northern Railway and U. S. route 91, both connecting Helena and Great Falls, follow the valley through the mountains.

FIELD WORK

Field work was done during 3 months in each of the summers of 1939 and 1940 and 1½ months in the fall of 1941. W. C. Adelhelm, D. L. Everhart, T. W. Noon, M. Pollard, and the late J. R. White assisted the writer. Mapping was done on aerial photographs (1:20,000) from the Department of Agriculture's A. A. A. The data were replotted on county township maps and from these were reduced to the present scale.

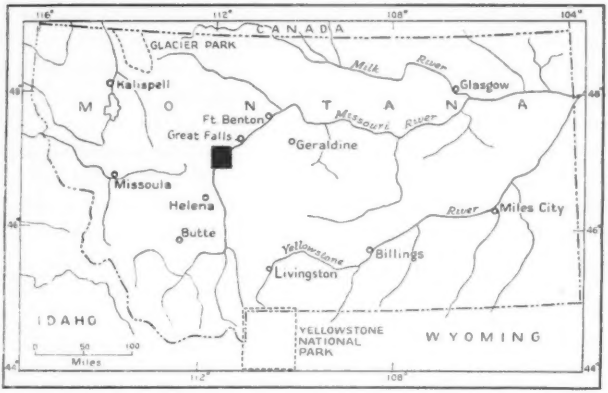

FIGURE 1.—*Map of Montana showing location of area*

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PREVIOUS WORK

This is the first study of the igneous rocks of the Big Belt Mountains, and the literature on them is scant. Lindgren (1886) first noted volcanic rocks at the northern edge of the range. Barnett (1917) and Fisher (1909) described the geology of the contiguous areas on the east and south in their investigation of the coal and water resources of the Great Falls district. Knopf (1913) and Pardee and Schrader (1933), in studying the Helena mining region, noted several salient structural features of the Big Belts. Calhoun (1908) and Alden (1932) reported on the physiography of the plains east of the mountains. Perry (1932) discussed the water supply of the town of Cascade. To the writer's knowledge, no other reports deal with any of the local geology.

PHYSIOGRAPHY

The most arresting feature of the drainage pattern of the Big Belts is the manner in which the Missouri River cuts directly through the mountains. Two of the main streams of the district, the Dearborn River and Flat Creek, enter the mountains from the plains to the north at an elevation of 3700 feet and have cut deep gorges through a part of the range whose peaks reach 6000 feet in elevation. These facts as well as the meandering courses of the main streams indicate considerable superposition of the drainage.

The literature (Alden, 1932) on western Montana indicates that in early Tertiary, probably Oligocene, time, damming of the Snake River drainage by crustal warping and lava flows initiated a series of large lakes, in some of which a few thousand feet of sediments accumulated. One of these lakes overflowed the divide of the Big Belt Range northeast of Helena and initiated the present course of the Missouri River. Much of the drainage of the headwaters of the Snake River was thus captured by the new stream, and the continental divide was shifted 50 to 100 miles westward.

The northeastward-meandering course of the Missouri through the Big Belts, in view of the fact that both topography and structure plunge northwestward, is probably ascribable to an Oligocene peneplain recognized by Willis (1902) at about 8000 feet in the Lewis and Livingston ranges and by Pardee and Schrader (1933) at about 7000 feet in the Big Belts.

Conspicuous features of the Rocky Mountain front and its adjacent plains are three dissected bench levels which Alden (1932) dates as Miocene to Pleistocene. Within the area of this report the Number 1 (Flaxville) bench is poorly developed and is represented, if at all, only by such igneous-capped mesas as Fort Shaw, Crown, and Square buttes, 4500 feet above sea level. Number 2 bench (Pliocene or early Pleistocene) is extensively developed in the plains northeast of Cascade, in pediments along the foot of Mount Harris, and in a series of benches along the Missouri River valley near Holter Dam at elevations of 3850 to 3900 feet. Number 3 bench (Pleistocene) is found near Holter Dam 3600 feet above sea level. In northern Montana glacial debris covers all these benches.

The effects of Illinoian and Wisconsin ice advances are not apparent in the Big Belts. A few boulders, 10 feet in diameter, near the mouth of West Hound Creek were possibly carried there by ice, but there is no evidence that this part of the range provided sufficient gathering ground for even small mountain glaciers.

Connected, however, with the Illinoian continental glaciation was the development of glacial Lake Great Falls bounded on the south, according to Alden (1932) and Calhoun (1906), by the Highwood, Little Belt, and Big Belt ranges. In the canyon of the Missouri, well-bedded yellow sands are now found, with a maximum preserved thickness of 150 feet, at elevations up to 3800 feet. No evidence as to the maximum depth of this glacial lake exists, but its surface may well have been at least as high as the 3900 feet assigned it by Alden.

As pointed out by Fisher (1909), well borings show that east of Cascade the Missouri now flows on a flood plain 270 feet above its rock-valley floor. Therefore extensive changes have occurred in local base level, probably since Pleistocene time.

STRATIGRAPHY

REGIONAL DISTRIBUTION

The northern part of the Big Belt Range is underlain chiefly by upper Cretaceous volcanics, 3200 feet thick, extruded unconformably upon upper Cretaceous sedi-

AGE		FORMATION	LITHOLOGY	THICKNESS	SYMBOL	COLUMNAR SECTION
QUATERNARY	Recent and Pleistocene	Alluvium	Alluvial fill	0-270	Qal	
		Great Falls lake sands	Fine-grained well-bedded yellow sands; found only in river valleys	0-250	Qgfs	
		UNCONFORMITY				
CRETACEOUS	Upper	Adel Mountain volcanics	Trachybasalt and basalt agglomerates, conglomerates, and flows, with numerous intrusives; analcime and augite trachybasalts in upper measures; contains plant fossils in upper conglomerates; entire succession strongly zeolitized	3200	Kamv	
		UNCONFORMITY				
		Two Medicine formation	Graywacke, shale, and sandstone, with drab gray and green graywackes abundant	1000	Ktm	
		Eagle sandstone	Massive buff concretionary sandstone; coal 150 feet above base	200	Kes	
		Colorado shale	Gray to black fissile shale; gray sandstone members near base and top	1500	Kc	
	Lower	Kootenai formation	Red sandy shale, with buff sandstone near base; coal 60 feet above base	570	Kk	
JURASSIC	Upper?	Ellis formation	Sandstone, grading upward into variegated shale, limestone, and sandstone	350	Je	
		UNCONFORMITY				
MISSISSIPPIAN	Middle and Upper	Big Snowy group	Shaly limestone, reddish toward base; green limy shale toward top	500	Cbs	
		Madison limestone	Massive gray limestone, weathering white; somewhat siliceous	1000	Cm	
CAMBRIAN DEVONIAN		Unexposed	Chiefly quartzite shale and limestone	2300?	?-D	
ALGONKIAN		Belt series including Spokane shale	Red and green argillite, with some quartzite and limestone; mostly unexposed	12000	As	

FIGURE 2.—*Columnar section of northern Big Belt Range*

mentary rocks and, with them, folded into northwesterly trending anticlines and synclines. Their western limit is marked by an imbricate zone which includes the great Lewis overthrust. East of the volcanics the gentle Sweetgrass arch extends northwesterly from the Little Belt Mountains to Belly River, Alberta.

Figure 2 shows the main lithologic features of the rocks. Because Fisher (1909a; 1909b) and Barnett (1917) described the stratigraphy of most of these formations in the near-by Great Falls area, much of the detail has been omitted.

SPOKANE SHALE

Of the seven Algonkian formations into which Walcott (1899) divided the Belt series of the Little Belt Range, the fifth from the base is a succession of alternating red and green argillites 4000 feet thick. It crops out 1 mile west of Wolf Creek, where

it is brought up along the Lewis overthrust and overrides the upper Cretaceous Adel Mountain volcanics.

CAMBRIAN AND DEVONIAN FORMATIONS

Approximately 2300 feet of Cambrian and Devonian quartzite, shale, and limestone underlies the Madison limestone but is nowhere exposed. Farther south these formations constitute, with the Belt series, the core of the Big Belt anticlinal uplift.

MADISON LIMESTONE

This Mississippian formation, a succession of massive gray limestones, weathering white and carrying intercalated shaly members in their lower half, has not been studied in detail where, as on the western side of the district, it becomes involved in the complicated structures. The probable total thickness of the formation here is 1100 feet (Sloss, 1942).

Sporadically abundant corals, crinoids, and brachiopods occur in the Madison. The presence of cup corals and the tendency of the limestone to be a cliff former are perhaps its most diagnostic features.

BIG SNOWY GROUP

The strata overlying the Madison limestone in this part of Montana have been considered Pennsylvanian and a part of the Quadrant formation. Their identity as upper Mississippian and the subdivision of the group into the Kibbey, Otter, and Heath formations have, however, been well established by Scott (1935).

The stratigraphy at the northern end of the Big Belt Range is complicated by an angular unconformity between the Big Snowy group and the overlying Jurassic Ellis formation and by the fact that the group is much folded and faulted. Along the Missouri River the Big Snowy consists of approximately 340 feet of alternating white limestones, red limy shales, and buff sandstones, the probable equivalent of the basal Kibbey formation, and 160 feet of limestone and green limy shale, probably the Otter formation. The Heath formation was not recognized at the top of the section.

Several genera of brachiopods are abundant within the Big Snowy and have been listed by Scott (1935).

ELLIS FORMATION

This formation consists essentially of a basal limestone 20 feet thick overlain by 60 feet of buff sandstone and 220 feet of variegated, thin-bedded limestones, sandstones, and shales. The upper 150 feet was considered by Fisher (1909a) and Barnett (1917) to represent the Morrison formation and to be lower Cretaceous, but the Morrison, like the Quadrant, is not developed in this part of Montana.

The crinoid *Astarte* sp. and the brachiopod *Rhynchonella gnathophora* are the most abundant and diagnostic fossils in the Ellis and are probably upper Jurassic.

The angular unconformity between the Ellis and the underlying Big Snowy group was nowhere observed, but it has been clearly demonstrated by Barnett (1917) in the Hound Creek district.

KOOTENAI FORMATION

This lower Cretaceous series of vivid vermilion shales and sandstones overlies the Ellis conformably. Sixty feet above its base a very prominent coarse buff sandstone is an excellent horizon marker. Immediately beneath are valuable coal beds which south of Great Falls, reach a thickness of 20 feet, but, where the Kootenai is exposed along the Missouri River, they are too shaly to mine.

COLORADO SHALE

This, the most widespread sedimentary formation of the district, is an assemblage of drab gray shales with interbedded sandstones which become important only near the top and base. It is 1500 feet thick, upper Cretaceous, and is important because a bentonite member near its center and a rhyolite tuff 200 feet above its base indicate the beginning of upper Cretaceous volcanism in this part of the Cordilleran geosyncline. Beside its distinctive appearance it is easily recognizable through such fossils as *Scaphites ventricosus* and *Inoceramus labiatus*.

EAGLE SANDSTONE

This buff, massive concretionary sandstone, 90 to 200 feet thick, underlies many bluffs west of Cascade and is the formation into which the large sills of the Big Belts have been injected. Its strata grade into the Colorado shale, and the contact is placed arbitrarily at that horizon where the sandstone constitutes the major, and the shale the minor, part of the sedimentary column.

From 10 to 40 feet below the top a bituminous coal member, 6 feet thick, is mined at several places northwest of Cascade. The nearest occurrence of coal in this formation known to the author outside the Big Belts is in the Judith Basin, 40 miles southeast of Great Falls (Bowen, 1915).

TWO MEDICINE FORMATION

The beds conformably above the Eagle sandstone consist of green to gray shales, sandstones, graywackes, and tuffs here correlated with the Two Medicine formation of northwestern Montana (Stebinger, 1914). As initially deposited the formation was over 1000 feet thick, but folding and erosion before the main extrusions of the Big Belt volcanic rocks make lateral and vertical correlation difficult.

Near St. Peter's Mission the basal part carries such fossils as *Trachodon* sp., *Unio* sp., *Siliqua* sp., *Melania whiteavesi*, and *Ostrea subtrigonalis*. Dinosaur fragments and fossilized evergreen wood are conspicuous.

The upper part exposed directly under the lavas of Cobern Mountain has a coquina consisting of *Ostrea subtrigonalis* or *Ostrea glabra*, and *Melania whiteavesi*. L. L. Sloss (personal communication) says: "The age is very probably Judith River, but as similar forms occur in the Bearpaw and Fox Hills it might be safer to assign the assemblage to the Two Medicine formation."

This formation, as developed in the Big Belts, is considered the equivalent not only of the type Two Medicine strata (Stebinger, 1914) but also of that part of the Liv-

ingston formation of southwestern Montana which Stone and Calvert (1910) have demonstrated to grade into the Judith River and Claggett beds. Large upper Cretaceous graywacke roof pendants (Knopf, 1913) in the Boulder batholith are also probable stratigraphic equivalents of the Two Medicine.

The graywacke minerals are chiefly andesine, augite, quartz, hornblende, and the very distinctive high optic-angle sphene, keilhauite. Sphene is never present in the volcanic rocks and their related intrusives in the Big Belts. This fact and the marked difference of the sodic plagioclase of the Two Medicine from the highly zoned calcic plagioclase of the volcanics make it extremely improbable that the formation is of local derivation.

ADEL MOUNTAIN VOLCANICS

This 3200-foot thick sequence of upper Cretaceous potash-rich volcanics and related intrusives unconformably overlies all the Cretaceous strata in this area. Its petrology, age, and distribution will be discussed later.

GREAT FALLS LAKE SANDS

These well-bedded Pleistocene sands rest unconformably upon the older formations and attain a maximum preserved thickness of 150 feet at Craig. Areally they are unimportant.

ALLUVIUM

The valley fill of the river bottom is locally 275 feet thick, as shown by well borings in the Missouri River valley at Ulm. It is both Pleistocene and Recent.

STRUCTURAL GEOLOGY

REGIONAL SETTING

Structurally the Big Belt Range is a complicated anticlinal body, broken by numerous thrusts, which lies west of the gentle arch of the Little Belt Range. Its core is underlain by a classic Algonkian section 12,000 feet thick (Walcott, 1899), and its western side is marked by a series of great thrusts including the Lewis overthrust. In the latitude which marks the southern limit of Plate 1 the Big Belt anticlinorium plunges northward, thereby directly affecting the structure.

The volcanics of the northern part of the range are folded and warped, the intensity of folding increasing westward, where a fault throws Paleozoic strata over the volcanics, and a block 5 miles wide west of this fault becomes a great imbricate zone. Folding before the extrusion of the volcanics structurally complicated the older rocks, but, since the major diastrophism postdated the volcanism, the major axes of folding remain parallel.

LEWIS OVERTHRUST

This fault, the largest in the area, is 1 mile west of Wolf Creek, where it thrusts Spokane shale over Adel Mountain volcanics. The dip of the fault plane is low: the

strike averages N.20°W.; the stratigraphic throw approximates 10,200 feet. No data on the amount of overriding by the thrust sheet can be found.

Clapp (1932) has traced this fault from north of Glacier Park, through the western side of the Big Belt Range (where Paleozoic and Algonkian rocks are involved), to the vicinity of Three Forks, at about Lat. 46°N. In its southern extension it is known as the Lombard overthrust (Haynes, 1916).

CRAIG IMBRICATE ZONE

Between the Lewis overthrust on the west and the volcanics on the east nine thrusts have piled the Mississippian, Jurassic, and Cretaceous formations into a shingle-block pattern. These faults parallel the regional structural trend of N.40°W., are high-angle thrusts (generally from the west), and have a stratigraphic throw of approximately 500 feet and a net slip of 1500 feet. They die out in the folded strata to the southeast, so that along Cottonwood Creek only five are well defined, and the imbricate zone merges into the anticlinal structure of the main Big Belt uplift.

The easternmost fault, the Craig thrust, is the largest, with a maximum stratigraphic throw of 3750 feet and a maximum net slip of 7500 feet in sec. 30, T.15N., R.3W. Its magnitude decreases both to the northwest and the southeast, but the fault is traceable along the entire eastern side of the range and marks the eastern limit of the main structural uplift. The strike locally deviates 30° from its main trend of N.40°W. Its dip of 49°W. could be observed only where the fault is crossed by the Missouri River. As the strike is so variable, the dip also is probably variable.

West of Craig, where the complicated anticline plunges abruptly northwestward at an angle of 50°, the faults exhibit sharp variations in strike and dip and appear to have been deformed. This is not unusual, for Clapp (1932) reports that high-angle thrust faults are among the earliest structural features developed within the folded rocks of northwestern Montana and that their deformation is common.

COBERN MOUNTAIN FAULTS

Cobern Mountain is a horstlike uplift east of a graben (Pl. 1). The fault planes are unobservable, and this structural interpretation is based on (1) the map plan of the lithologic units, and (2) the extreme improbability of an erosional unconformity of the magnitude necessary to produce the map plan.

The transverse fault 1½ miles north of Cobern Mountain brings the Two Medicine formation into contact with conglomerates of the upper part of the Adel Mountain volcanics. The throw is more than 900 feet The attitude of the fault plane could not be observed, but a light-colored monzonite dike was not perceptibly offset by it in ground plan. Hence the monzonite is postfaulting—a remote possibility—or the dike and the fault have a nearly vertical dip, or the net slip along the fault paralleled the intersection of the dike and fault planes.

Only two other transverse faults have been mapped within the Big Belts, one near the mouth of Hardy Creek in T.17N., R.2W., where the volcanics and Colorado shale are probably faulted against one another, and the other near the junction of the north and south forks of Sheep Creek, sec. 7, T.16N., R.1W. Like that north of Cobern

Mountain, both have their southeasterly side brought up. All these structures appear related to an axis of cross warping which corresponds with the Missouri River Valley and which will be discussed later. Clapp (1932), in his study of the structure of northwestern Montana, found many transverse faults, all later than the main thrust faulting.

FAULTS WITHIN THE VOLCANICS

The massive character and poor bedding in the volcanics, the difficulty of separating intrusives and extrusives, and the lack of horizon markers have made it impossible to distinguish faults in them.

In sec. 7, T.16N., R.2W., two normal faults trend N.20°W., one on the east bringing Colorado shale into contact with the volcanics, and that on the west bringing together oppositely dipping lavas. Good exposures on sedimentary layers within the volcanics between the two faults indicate a number of small normal faults, with the eastward side rising steplike. An uplift along an anticlinal axis in the sediments beneath is doubtless responsible for this phenomenon.

A fault near the head of Encampment Creek in sec. 30, T.15N., R.1E., brings Kootenai beds into contact with Colorado shale, but where the fault enters the volcanics it cannot be traced. Possibly this fault is pre-volcanic, but, if so, it is the only one thus recorded.

FOLDS

The pre-volcanic structure was that of a dome whose center lay south of the mapped area. Near West Hound Creek at least 2400 feet of sediments was stripped from the northern slope of the dome before the volcanism. The structural relief of the broad upwarp must have far exceeded 2400 feet and possibly equalled the 3500 feet it has today. Much of the broad upwarping of the Little and Big Belt Ranges may have been produced at this time, and the later diastrophism may have corrugated and faulted these earlier master structures.

The maximum recorded angular divergence of the underlying sediments to the overlying volcanics is in sec. 11, T.16N., R.4W., where the Two Medicine strata dip 30°SE., and the overlying lavas are flat.

The five major anticlines shown within the volcanic rocks are poorly defined, and the folding is less regular than indicated on Plate 1. The westernmost three anticlinal axes can be traced fairly accurately across the entire area. The northeastern end of the volcanic field seems, however, to be devoid of any northwesterly-trending fold axes.

Near the mouth of Hardy Creek the volcanics dip up to 50°SE. The folds southeast and northwest of Craig plunge toward the Missouri valley bottom, and the imbricate block plunges northwestward. At the northern end of the volcanic field the contact of the lavas and sediments is at an elevation of 3700 to 3900 feet, and at the southwestern edge approximately 5300 feet. In the Missouri River valley, less than 3400 feet above sea level, the sedimentary rocks are covered. Therefore, a definite structural trough must have developed at right angles to the regional structural trend

in these folded volcanics, and its presence accounts in part for the course of the river.

The cause of this trough is not apparent and seems unrelated to late deformations across the structure, the presence of igneous bodies, or pre-volcanic structures. Deformations such as those in the southern Absaroka Mountains (Rouse, 1940), where the volcanics are folded and their underlying sedimentary rocks relatively undisturbed, are clearly not repeated within the Big Belts. Isostatic settling of the volcanic mass and its gradual rise during the Tertiary may be responsible not only for the cross warp but also for the transverse faults.

The minor folds within the Craig imbricate block have wave lengths, depending on their position, of from 200 to 500 yards, and overturning is common near the Craig thrust. Well-developed drag folds occur along this fault in all formations except the Madison limestone.

STRUCTURAL RELATIONS OF IGNEOUS ROCKS

INTRODUCTION

The chief intrusive members of the Adel Mountain volcanics are trachybasalts, syenogabbros, and monzonites which occupy approximately $\frac{1}{20}$ of the volume of the entire formation. Many trachybasalt intrusives could not be differentiated from the flows, and so are not indicated as hypabyssal bodies. Plate 1, therefore, minimizes their quantitative importance.

So far as known all the igneous rocks were intruded before the main diastrophism. The best evidence is the faulted syenogabbro and hornblende monzonite dikes and the folded volcanic conglomerates which contain fragments of all but the very latest intrusives.

DIKES

The most pronounced feature of the northern Big Belts is the radial arrangement of the many dikes with respect to the Three Sisters stock. The width of these dikes is from 8 to 50 feet, with 15 or 20 the average. Their attitude is dependent on bedrock and structure. Close to the main stock, and in the sedimentary rocks which they intrude, they are vertical, but within the volcanics a few miles from the intrusive center their attitude and trend are chaotic.

The longest mapped dike trends northward for over 10 miles toward Shaw Butte and is exposed throughout its length. It ends 1 mile south of the butte, but its projected strike is continued by two or three dikes of the same composition. Most of the dikes along the same line of strike are probably connected, but it is very difficult to prove this within the volcanics because the dikes are not well exposed by differential weathering.

Beside the radial arrangement, a well-marked north-south alignment is also apparent, possibly due to intrusion parallel to the pre-volcanic axes of folding.

A tendency for one rock type to be concentrated in a swarm is also very noticeable. On the divide between the headwaters of Stickney and Wegener creeks, for example, there are 15 dikes of a monzonitic type found nowhere else in the area, and this is by

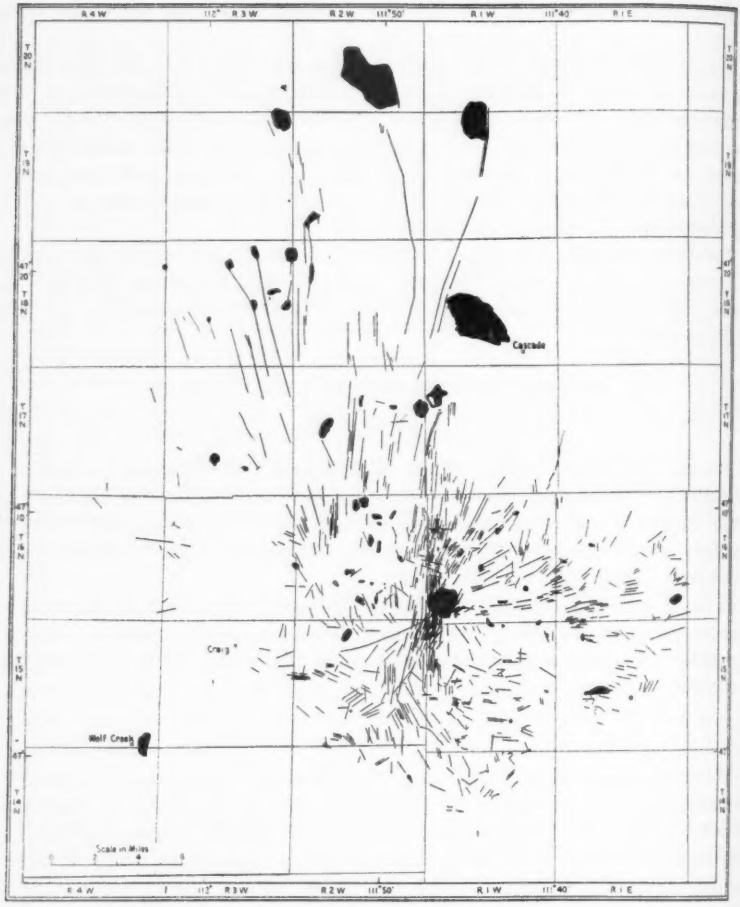

FIGURE 3.—*Map of intrusive rocks of northern Big Belt Range*

no means a singular example. These localized swarms are due to the presence of a vertical feeder of the same composition or to directed horizontal intrusion along fractures connecting with the main magma chamber.

Hexagonal jointing normal to the walls is well developed in many of the dikes, but especially in the more mafic differentiates. The only structural abnormality is represented by Fishback Butte in sec. 7, T.18N., R.2W., where a syenogabbro dike has spread its walls to a width of 300 feet and has developed a huge lens-shaped body a mile long.

The direction of intrusion and the cause of the radial pattern are problems to which this area can contribute little new information. The ideas of Richey and Thomas

(1930) and of Iddings (1914) that radial fractures and dike systems are due to excessive magmatic pressure acting vertically upward on a relatively thin cover have met with no serious disqualifying objections and are applicable to the structures observable in the Big Belts. The reason for the localization of the Three Sisters volcanic vent is unknown.

The direction from which the dikes were intruded—*i.e.*, vertically from below, or horizontally from a central magma chamber is not certain. Parsons (1939), in the Sunlight area of Wyoming, found that radial dikes around a volcanic vent cut the surrounding lavas, but not the Paleozoic rocks exposed by erosion beneath them. This he considered evidence for horizontal injection of the dike material from the central magma vent.

Within the Big Belts no dikes cut the Paleozoic strata, but they are extremely common in the upper part of the stratigraphic column. Clapp (1932) made a similar generalization for all of northwestern Montana. These facts, in addition to the evidence that the Eagle sandstone probably diverts some dikes into sills because of its great brittleness, are reconcilable with the proposal that the dikes of the Big Belts have spread laterally from the Three Sisters stock rather than vertically from below. If the intrusion had been upward from a batholith, the area of that batholith would have exceeded 800 square miles. Aside from the dikes themselves and the widespread zeolitization of all the igneous rocks, there is no surface indication of the existence of such a large intrusive.

None of the Big Belt dikes showed any preferred orientation of the augite or plagioclase phenocrysts.

SILLS

Of the three major rock types—trachybasalts, syenogabbros, and monzonites—only the first two form sills. The trachybasalts are confined to the volcanics, and the syenogabbros to the sedimentary rocks. This striking distribution is probably due to the fact that, when the syenogabbros were irrupted, the volcanics were too well consolidated to allow concordant intrusions within them, whereas the trachybasalts were intruded with some of the volcanic extrusions.

The great sills of the Big Belts, such as Fort Shaw, Cascade, Crown, and Square Buttes, have apparently been fed by syenogabbro dikes which head across the plains from the Three Sisters stock. Evidence for this is that some dikes bend over into the sills, but no dikes cut the sills or extend beyond them.

The jointing of the sills was initiated by a series of fractures parallel to the walls of the body. With continued cooling and differentiation of the syenogabbro, syenitic auto-differentiates were injected into the early joints. After solidification of the syenite, now from 4 to 6 per cent of the total rock volume of the sills, columnar joints formed normal to the retaining roof rocks.

A perplexing feature of several of the sills, but most notably of Square Butte, is a horizontal banding consisting, in that case, of 28 syenogabbro layers, each 8 to 10 feet thick, and each alternate one being darker, denser, and more resistant to erosion than the intermediate ones. The resultant spectacular stratification on the steep butte walls is continuous over the entire sill, and the structure is analogous to that

of a deck of cards (Pl. 2, fig. 1). Heavy-mineral and specific-gravity determinations made on some of these bands are shown in Table 1. The specific-gravity determinations are reproducible to ±0.02.

The failure of the specific-gravity and heavy-mineral determinations to agree more closely is due to extensive zeolitization of the rocks. The concentration of the

TABLE 1.—*Specific gravity and heavy mineral determinations, Square Butte sill*

Sample	Locality	Weight % heavies	Specific gravity
C.159	Dike; heads toward Square Butte	33.4	2.69
C.176	Dike; southeast side of Square Butte	45.3	2.70
C.174	Light band; Square Butte	36.4	2.64
C.179c	Light band; Square Butte	34.4	2.63
C.376	Light band; Shaw Butte	38.5	2.62
C.166	Dark band; Square Butte	39.8	2.72
C.173	Dark band; Square Butte	36.9	2.73
C.175	Dark band; Square Butte	44.3	2.74
C.178b	Dark band; Square Butte	41.7	2.77
C.377	Dark band; Shaw Butte	40.6	2.70

TABLE 2.—*Specific gravities of rocks from the bottom of a dark layer to the top of a light one, Square Butte sill*

Specimen	Locality	Specific gravity	Feet from bottom
C.178a	Bottom of a dark layer	2.68	1
C.178b	Middle of dark layer	2.77	4
C.178c	Top of dark layer	2.73	7
C.179a	Bottom of light layer	2.69	9
C.179b	Middle of light layer (too friable to sample)		12
C.179c	Top of light layer	2.63	15

heavy minerals in the dark bands is, however, clearly demonstrated, and there is no noticeable decrease in the heavy-mineral content from the bottom to the top of the sill.

The contact between the dark and the light bands is gradational, as shown by determinations made on samples taken from the bottom of a dark to the top of a light band (Table 2).

Cascade and Fort Shaw Buttes exhibit less well-developed analogous structures. Crown Butte, however, shows only a general gravity decrease toward the center of the sill, which is probably due to the effects of zeolitization.

Because of the hydrothermal alteration of all the sills, and because their roofs and walls have been stripped away, we can only guess at the magmatic processes producing this banding. Obviously, theories which explain layering on the basis of the sinking of the floor of the intrusive are inapplicable here. Hess' (1938) proposal of (1) relative density differences between each of the two main constituents settling and (2) the disturbance of the normal state of quiescence by short epochs of mild but irregular turbulence may be suited to the present case. An equally plausible idea is

that 14 separate injections have occurred at Square Butte, each one differentiating into a light and a dark band. Still another theory is that some such process as that which produces Liesegang rings also produced the banding, but there is no direct information on how this mechanism could operate in an igneous body. Unfortunately no theory can be directly proven, and the proposed explanations, like those for almost all banded igneous structures, are unsatisfactory.

LACCOLITHS

The syenogabbro laccoliths of the Big Belt Mountains occur in T.18N., R.3W., where there are five such bodies in the Two Medicine formation. This interesting correlation of the Two Medicine formation with the laccoliths, and of the Eagle sandstone with the sills, reflects the brittleness of the sandstone and the plasticity of the Two Medicine shales and graywackes.

These intrusives are classified as laccoliths chiefly on the form of the outcrop, the petrographic character of the rocks, and their jointing. Jointing similar to the horizontal jointing of the sills overlies all the buttes, and a well-defined columnar jointing is developed normal to the first jointing system, and presumably normal to the now-missing roof rocks. In no case, however, is the floor of the intrusive well exposed, nor can it be demonstrated, for lack of outcrops, that the sediments once arched over the roofs of these intrusives.

Dike feeders are present for all these intrusives, just as for the sills. Of particular interest are Lion's Mane Butte and the butte 2 miles southeast of it, both of which seem to have been fed from the same syenogabbro dike.

The trend of magmatic differentiation cannot be studied within the laccoliths because erosion has nowhere revealed good cross sections.

PLUGS AND CHONOLITHS

The Three Sisters composite stock, a more or less circular plug 1½ miles in diameter and composed of monzonites, syenogabbros, and trachybasalts, is considered the source of most of the volcanics of the Big Belt Mountains. The evidence for calling the stock an ancient volcanic throat is (1) a radiating composite dike system, (2) the great thickness of volcanics near by, (3) the slight epidotization peripheral to the stock, (4) the trachybasalt breccia dikes contiguous to the intrusive, and (5) its elevation, which is greater than that of every other peak except Adel Mountain. These criteria accord well with those found by Parsons (1939) and Rouse (1937) for similar structures in the Absaroka Mountains.

Few of the other abundant stocks and chonoliths in the mountains were extrusive vents. All the stocks are small and irregular.

No high-temperature metamorphism is found near the intrusives, and there has been no recrystallization or introduction of new material into the wall rock.

PETROGRAPHY

FIELD RELATIONS

The Adel Mountain volcanics occupy approximately 330 square miles of the northern Big Belts, exclusive of 20 square miles occupied by outlying intrusives and a strip

of volcanics 2 miles wide extending northwest and southeast of Wolf Creek. Within the orthoclase basalt series, 3200 feet thick, no horizon markers could be found, nor any regular stratigraphy worked out.

The volcanics consist of breccias, agglomerates, conglomerates, tuffs, flows, and graywackes. The typical rock is a flow breccia of abundant subrounded fragments (chiefly trachybasalt types) having diameters up to 12 feet. Their great diversity suggests that these boulders came from the vent itself rather than that they were picked up on the surface by the lavas. Volcanic conglomerates are areally important only near Craig. Massive flows, distinguished with great difficulty from the sills, are common in the upper part of the series.

In the following descriptions, the rock types are discussed from oldest to youngest, regardless of their mode of formation.

PLAGIOCLASE CALCIRHYOLITE

This rock forms a plug at Wolf Creek and a massive, poorly bedded flow 165 feet thick northwest and southeast of it. It directly overlies the Two Medicine formation and underlies a part of the trachybasalts; hence it is probably the oldest Adel Mountain volcanic.

The 3-millimeter plagioclase phenocrysts are calcic labradorite. Smaller andesine phenocrysts are enclosed in the barium orthoclase of the groundmass. Ten per cent quartz and a few per cent each of augite, apatite, magnetite, and nontronite complete the mineralogical assemblage.

QUARTZ LATITE

A series of quartz latites examined a few miles northwest of Wolf Creek is probably stratigraphically above the calcirhyolites. Whether or not these flows are younger than the main trachybasalts of the range is uncertain, but clearly they do not, as Pardee and Schrader (1933) state, postdate the Lewis overthrust.

The quartz latites are composed of phenocrysts ranging from andesine to labradorite, a few rare biotite and augite phenocrysts, and a cryptocrystalline groundmass of quartz, tridymite, oligoclase, and orthoclase. Apatite, zircon, and magnetite are the accessories.

The calcirhyolite and quartz latite are confined to the area near Wolf Creek. Rhyolite tuffs are found, however, within the Two Medicine formation.

TRACHYBASALTS

These, the most important extrusive rocks, are potash-rich volcanics similar to the absarokite-shoshonite-banakite series first described by Iddings (1899) in the Absaroka Range. They are predominantly purple and gray breccias and agglomerates (Pl. 2, fig. 2), but related massive black flows and intrusives are common. Some fragments in the agglomerate are as much as 12 feet in diameter (Pl. 2, fig. 3).

The typical trachybasalt contains from 10 to 15 per cent modal orthoclase which moulds around highly zoned calcic labradorite phenocrysts. The augite phenocrysts are commonly 5 mm. long. Their β index of refraction, and that of all the Adel

FIGURE 1.—INTRUSIVE INTO BANDED STRUCTURE OF SQUARE BUTTE

FIGURE 2.—POORLY BEDDED BRECCIA, SHEEP CREEK

FIGURE 3.—TYPICAL SUBROUNDED FRAGMENT IN BRECCIA, SHEEP CREEK

IGNEOUS STRUCTURES

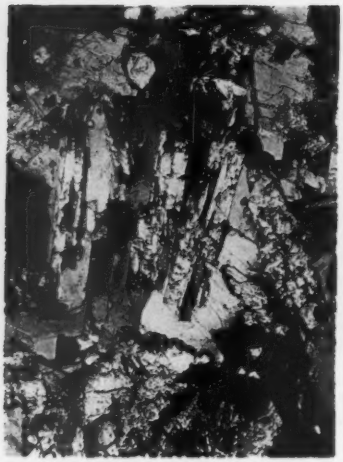

FIGURE 1.—SYENOGABBRO
Calcic labradorite enclosed in orthoclase.
Crossed nicols. X48

FIGURE 2.—SYENITIC DIFFERENTIATE OF SYENOGABBRO
Biotite oriented on hornblende it replaces.
Plane polarized light. X60

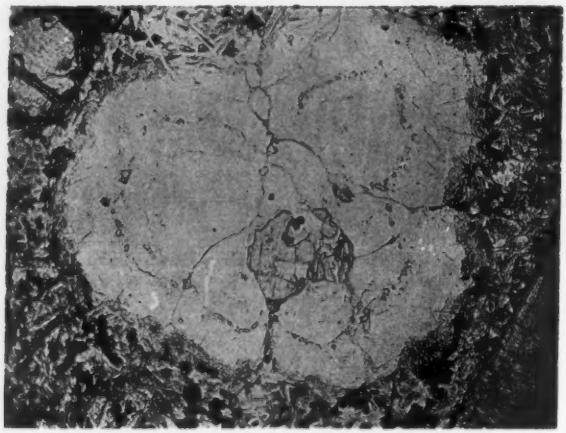

FIGURE 3.—INTERGROWTH OF FIVE ANALCIME PHENOCRYSTS
Groundmass consists of fresh labradorite. Plane polarized light. X60

PHOTOMICROGRAPHS

Mountain volcanics, ranges from 1.695 to 1.711. This small range and the association of orthoclase with calcic plagioclase are the most diagnostic features of the entire petrographic province.

The olivine, where unaltered to iddingsite, chlorite, or nontronite, is surprisingly high in iron and contains from 25 to 49 per cent of the fayalite molecule. This is totally unexpectable on the basis of Bowen's (1935) MgO-FeO-SiO_2 diagram and probably reflects the high iron content of the parent magma. The concentration of the iron in the olivine in preference to the pyroxene is unexplained, unless it be due to the earlier period of formation of the olivine in the iron-rich environment.

Hornblende and biotite are rare. Large apatite and magnetite crystals are abundant accessories. The variable color of the volcanics is due to varying quantities and types of hydrous iron oxides and hematite.

Most of the groundmass is replaced by the zeolites, which are the most important minerals in the area. As all the igneous rocks have been affected by this widespread replacement further discussion will be postponed until all the rocks have been separately described.

ANALCIME TRACHYBASALT AND AUGITE TRACHYBASALT

Analcime trachybasalt and augite trachybasalt intrusives and flows are mineralogically identical with the common trachybasalt extrusives, except that the analcime trachybasalt contains from 5 to 15 per cent analcime phenocrysts 5 millimeters or less in diameter (Pl. 3, fig. 3). Megascopically these rocks are distinguished by their dark black matrix and orange-colored weathered surface. Both have characteristically large augite phenocrysts.

They are almost contemporaneous and appeared after the bulk of the volcanics had been extruded. Some of the augite trachybasalts are probably very young, and related types certainly postdate all but one of the very late differentiates.

BIOTITE PYROXENITE

Coarsely crystalline biotite pyroxenite forms inclusions up to a foot in diameter in all intrusive types within the Three Sisters stock. It is composed of equal quantities of biotite and augite, with a very small amount of interstitial magnetite, apatite, analcime, stilbite (?), and andesine. This is considered the earliest of the strictly hypabyssal rocks of the Big Belt sequence. Its distinct contact with compositionally different host rocks makes it fairly certain that the biotite pyroxenite is a distinct petrologic unit.

SYENOGABBRO

Syenogabbro, the chief intrusive type of the mountains, is gray with augite phenocrysts a centimeter long in a matrix of plagioclase, strongly zoned and averaging labradorite, which are imbedded in barium sanidine (Pl. 3, fig. 1). Its chilled dike and sill contents are surficially like some of the trachybasalts, although its more felsic varieties are light-colored and grade into syenite.

The syenogabbros have the same minerals as the trachybasalts, but orthoclase (barium sanidine) is as abundant as plagioclase, and the mafic minerals are fewer. Optical and chemical properties of all the minerals are similar.

In large intrusive bodies the syenogabbro generated, by crystallization, a residual syenitic liquor which was injected into the joints of the cooling intrusive and solidifies in a sheet about an inch thick. This syenite is white, medium-grained, with an equigranular texture, and contains a few hornblende phenocrysts. Microcline, albite, aegirite-augite, biotite, and the accessories apatite and magnetite are the other minerals. The biotite in part replaces and grows about the hornblende (Pl. 3, fig. 2). Its genesis from the syenogabbro can be proven at several places by gradational segregations.

HORNBLENDE MONZONITE

This rock forms several small stocks (Finnegan Mountain, Sawmill Peak, and a plug at the mouth of Sullivan Creek) and many dikes, but no sills or laccoliths. Its dikes cut syenogabbros but are intruded by augite calcimonzonite porphyry, biotite-augite diorite, and monzonite.

The slender amphibole phenocrysts are basaltic hornblende. The strongly zoned plagioclase (An_{40}) is more abundant than the orthoclase. Green biotite, quartz (rare), magnetite, chlorite, apatite, and nontronite are the remaining rock minerals.

AUGITE CALCIMONZONITE PORPHYRY

This rock is distinguished megascopically by its large augite phenocrysts in a light-gray matrix. Smaller labradorite phenocrysts also appear, but not so conspicuously. The groundmass feldspar is oligoclase and orthoclase. Biotite and the accessories are the only other important primary minerals.

The calcimonzonite porphyry is intruded by late monzonite dikes and cuts hornblende monzonite intrusives.

QUARTZ MONZONITE PORPHYRY

A light-gray holocrystalline rock with orange to white plagioclase phenocrysts in an aphanitic matrix is found as fragments in the volcanic conglomerates west of Craig and as dikes cutting 3000 feet of trachybasalt near the Three Sisters stock. The phenocrysts are zoned oligoclase, with minor inclusions of a more calcic, but indeterminate, plagioclase. Cryptocrystalline orthoclase and about 10 per cent quartz constitute the matrix. Biotite, hornblende, augite, zircon, and magnetite, although present, are quantitatively unimportant.

MONZONITE

One of the younger intrusives of the igneous complex is a light-gray, medium-grained, subporphyritic monzonite in which small augite crystals are conspicuous on the weathered surface. This is younger than all the rock types heretofore described and has nowhere been observed cut by another intrusive. It does occur, however, as boulders in the volcanic conglomerates west of Craig. A large portion of the Three Sisters

stock, many dikes, and the Antelope Mountain stock are made up of this monzonite.

Augite is the chief mafic mineral, but biotite and hornblende are also present. The strongly zoned plagioclase is calcic andesine in a matrix of orthoclase. Apatite, magnetite, and groundmass augite are the other minerals.

SANIDINE MONZONITE

This rock is conspicuous because of 10 per cent of barium sanidine phenocrysts set in a light-gray, fine-grained, pilotaxitic matrix which contains smaller phenocrysts of strongly zoned andesine. Orthoclase, apatite, magnetite, augite, biotite, and hornblende form the groundmass. Dikes of sanidine monzonite, though rare, everywhere truncate the rocks with which they are in contact.

Sanidine phenocrysts in a rock whose normative feldspar approaches that of this rock is not explicable by Bowen's (1928, p. 227–233) theoretical equilibrium diagram for the feldspars. On Larsen's diagram (1938) constructed from the analyzed feldspar phenocrysts of the San Juan lavas, the sanidine could easily be in equilibrium not only with the magma but with the plagioclase.

BIOTITE-AUGITE DIORITE

A reversal in the normal progression from gabbro to quartz monzonite is represented by a black augite porphyry in which biotite is the only mineral visible in the aphanitic groundmass. This rock truncates hornblende monzonites in the form of dikes, but its relations beyond this point are unknown.

Zeolitization has so altered the primary minerals that except for the labradorite phenocrysts it is impossible to tell the composition of the groundmass plagioclase or the quantity of orthoclase. Apatite and magnetite are unaltered.

VOLCANIC CONGLOMERATE

Occupying a total area of approximately 25 square miles near Craig a series of well-bedded volcanic conglomerates contains rounded cobbles of almost all the previously described dike types. This portion of the volcanics is, then, younger than all the major intrusives. The well-bedded matrix is both latitic and trachybasaltic, but most of the matrices studied were too greatly zeolitized to be properly identified. Although large portions of the conglomerate are unquestionably detrital, some may be the result of renewed explosive activity.

ORTHOCLASE DIABASE

An orthoclase diabase which differs only slightly from the rocks previously described as augite trachybasalt intrudes the volcanics at Craig and contains labradorite, interstitial orthoclase, olivine, augite, magnetite, and apatite. The only perceptible mineralogical deviation from the earlier augite trachybasalt is that the pyroxenes are smaller in the orthoclase diabase and olivine is more abundant and higher in the iron molecule.

LIGHT-COLORED MONZONITE

This light-gray, fine-grained rock is distinguished chiefly by its saccharoidal texture and lack of phenocrysts. The monzonite cuts the conglomerate at Craig and is found only here and at Cobern Mountain. Presumably it is the youngest igneous rock of the mountains.

Orthoclase and oligoclase are present in the proportion of 2:1. Hornblende, augite, apatite, and zeolites are the remaining minerals.

ZEOLITIZATION

Aside from its high potash content the most striking feature of the igneous sequence is the degree and intensity of zeolitization of all the rocks. Broadly speaking the zeolites occur (1) as cross-cutting veinlets, (2) as amygdules, and (3) as replacements of the silicates. This last and most important type is not commonly recognized megascopically, and most of the zeolitized rocks appear fresh except for the stony appearance of their matrices.

Analcime, mesolite, thomsonite, stilbite, heulandite, apophyllite, natrolite, scolecite, and leonhardite are, in order of abundance, the important zeolites. Calcite is an almost constant associate. The most abundant alteration product is weakly anisotropic and has a mean index of refraction of 1.495 to 1.500. This corresponds to no mineral species and is regarded as a complex lime, soda, potash, aluminum silicate related to faujasite, or a high potash and lime analcime.

Zeolites replace plagioclase and orthoclase indiscriminately, and the groundmass is recrystallized, in general, to two minerals—the so-called faujasite, and stilbite, thomsonite, or heulandite. Augite, apatite, and magnetite are unaltered, but, if calcite is introduced, it preferentially replaces augite. Olivine changes to iddingsite or nontronite, but this reaction is probably deuteric.

No well-marked paragenetic sequence exists among the zeolites, nor is there any apparent reason why one zeolite forms in preference to another. Specularite, kaolin, epidote, and zoisite are the only other minerals with the zeolites, the last two occurring, however, only close to the Three Sisters plug.

The chemical changes brought about by zeolitization are insignificant, some water being added and some iron oxidized. This is shown by the two analyses in Table 3, both of rocks from the Square Butte syenogabbro sill. The first, C.188, is from the fresh rock found at the top of the sill; the second, C.185, from an altered dark band.

Zeolitization has affected nearly all the rocks, and there appears to be no systematic distribution of the more intensely zeolitized rock about intrusive bodies. However, zeolitization is greatest in areas where the rocks are fractured and in porous rocks. The augite trachybasalts are a good example; massive flows are relatively fresh, and agglomerates and breccias are invariably much altered.

Two theories might explain this zeolitization: (1) Volcanics and intrusives alike have been altered chiefly by ground water which seeped into the rocks shortly after their eruption, and before they cooled to atmospheric temperatures. (2) Some of the water is magmatic and was concentrated residually by the crystallization of the silicates. The first hypothesis fails to explain the zeolite veins outside the volcanic

area or the sporadic distribution of the zeolitization; the second meets with difficulties in the distribution of the water, especially in such areas as Square Butte 20 miles from the intrusive center. Perhaps both processes working together have been ultimately responsible.

TABLE 3.—*Analyses of fresh and altered rocks, Square Butte*

	C.188	C.185	% loss	% gain
SiO_2	48.96	47.40	1.56	
TiO_2	0.61	0.58	0.03	
Al_2O_3	17.04	16.28	0.76	
Fe_2O_3	3.90	5.26		1.36
FeO	5.02	4.73	0.29	
MnO	0.12	0.08	0.04	
MgO	4.54	4.47	0.07	
CaO	8.56	8.16	0.40	
Na_2O	3.59	3.18	0.41	
K_2O	4.75	4.46	0.29	
H_2O^-	0.08	—	0.08	
H_2O^+	1.97	4.27		2.30
CO_2	none	none	none	
P_2O_5	0.75	0.69	0.06	
S	0.05	0.08		0.03
BaO	0.17	—	—	
SrO	0.04	0.04	0.00	
	100.15	99.68		

AGE OF VOLCANICS

The igneous sequence can be dated fairly definitely. The volcanics rest upon a surface eroded across the Two Medicine formation, and both intrusives and extrusives are involved in the Laramide orogeny. Assuming that the orogeny is late Cretaceous or early Eocene, the volcanics are late upper Cretaceous.

In collections of fossil plants from the Two Medicine formation exposed at the Narragang ranch 3 miles west of Craig, and from conglomerates of the Adel Mountain volcanics 1 mile north of Craig, Erling Dorf found the following fossils:

A. From 3 miles west of Craig
   (1). *Quercus viburnifolia* Lesquereux (?)
   (2). *Cercidiphyllum ellipticum* (Newberry) Brown
   (3). *Dryophyllum subfalcatum* Lesquereux
   (4). *Viburnum hesperium* Knowlton

B. From 1 mile north of Craig
   (1). *Viburnum montanum* Knowlton
   (2). *Plantophyllum* sp., apparently new

He dates these fossils as late Cretaceous, probably Montanan.

On stratigraphic basis the Adel Mountain volcanics are apparently equivalent in age to some of the Livingston volcanics of southwestern Montana. This correlation is further strengthened by their petrology, both areas being characterized by potash-rich trachybasalts.

TABLE 4.—*Chemical analyses, norms, and modes of the Big Belt igneous rocks*

	C.185	C.188	C.223	C.224	C.229	C.244	21115
ANALYSES							
SiO_2	47.40	48.96	49.94	50.18	52.07	47.76	50.70
TiO_2	0.58	0.61	0.61	0.89	0.79	0.58	0.66
Al_2O_3	16.24	17.04	17.02	15.34	18.18	16.20	18.32
Fe_2O_3	5.26	3.90	6.05	6.37	4.17	3.98	4.62
FeO	4.73	5.02	3.87	4.44	4.73	6.45	3.01
MnO	0.08	0.12	0.13	0.12	0.09	0.16	0.09
MgO	4.47	4.54	4.25	4.11	3.33	4.90	4.09
CaO	8.16	8.56	9.22	8.62	7.58	8.60	8.68
Na_2O	3.18	3.59	2.81	4.06	3.57	3.20	3.33
K_2O	4.46	4.75	2.41	2.48	2.87	3.76	3.16
H_2O^-	—	0.08	1.11	0.36	0.65	0.31	0.60
H_2O^+	4.27	1.97	1.67	2.01	1.12	3.06	1.77
CO_2	none	none	none	none	none	none	0.12
P_2O_5	0.69	0.75	0.41	0.58	0.64	0.61	0.31
S	0.08	0.05	0.14	0.06	0.03	0.04	0.28
BaO	—	0.17	0.13	0.17	0.08	0.15	0.11
SrO	0.04	0.04	none	0.03	0.02	0.03	trace
Total	99.64	100.15	99.77	99.82	99.92	99.79	99.83
F. A. Gonyer, analyst.							
Position: $(\frac{1}{3}\ SiO_2 + K_2O - MgO - CaO - FeO)$	2.27	−0.92	−4.20	−4.21	0.69	−4.92	−0.70
NORMS							
Q			2.94		0.30		
or	26.69	28.36	14.46	15.01	17.24	22.24	18.90
ab	13.62	14.15	22.53	31.96	29.87	15.20	27.77
an	16.40	16.12	27.24	15.85	25.02	18.63	25.58
ne	7.38	8.80		1.42		6.53	
hy			5.69		8.60		
di	15.53	14.12	12.47	18.62	7.60	16.97	12.40
ol	6.50	7.07		2.12		8.51	3.51
ap	1.68	2.36	1.01	1.30	1.34	1.34	0.67
mg	7.66	5.57	8.82	9.28	6.03	5.80	6.73
il	1.06	1.22	1.22	1.67	1.52	1.22	1.37
tn			0.28				0.56
Total	96.52	97.77	96.66	97.23	97.52	96.44	97.49
MODES							
orth	31.4	26.8		16.8	21.4	27.9	25.1
plag	33.8	32.8	54.8	39.7	49.7	32.1	41.0
aug			16.2		18.4	23.3	17.0
ol	28.1	31.3		25.7	4.5		9.1
bio			15.1			12.1	
ap							
mg	6.8	9.1	9.8	9.7	7.0	4.6	7.8
analcime				8.2			
groundmass			4.2				
Total	100.1	100.0	100.1	100.1	101.0	100.0	100.0
Anorthite content of groundmass plagioclase	60?	62	65	60	63	60	58

C.185 Altered syenogabbro, Square Butte.
C.188 Fresh syenogabbro, Square Butte.
C.223 Trachybasalt boulder in agglomerate, Hardy Creek.
C.224 Analcime trachybasalt flow, Hardy Creek.
C.229 Trachybasalt dike, Hardy Creek.
C.244 Syenogabbro plug, sec. 10, T.16 N., R. 2 W.
21115 Trachybasalt boulder in conglomerate, Craig.

PETROLOGY

CHEMICAL PETROLOGY

The seven analyses of the Big Belt suite (Table 4) represent all but the monzonitic types.

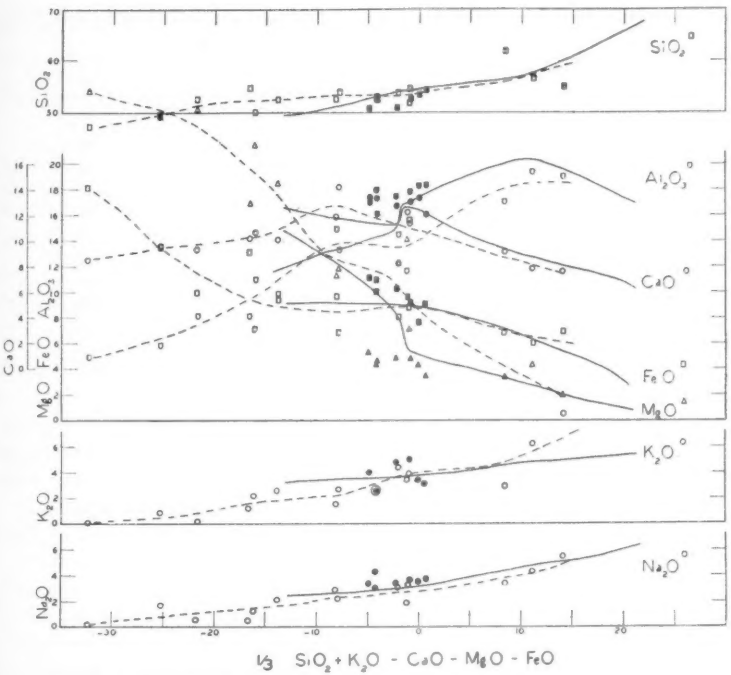

FIGURE 4.—*Variation diagram of Big Belt, Livingston, and Absaroka subprovinces*
Solid circles, squares, and triangles--Big Belt; open circles, squares, and triangles--Livingston; dotted curves--Livingston; solid curves--Absaroka.

These analyses have been plotted on a variation diagram (Fig. 4) constructed after the manner of Larsen (1938). With them are plotted the variation curves of the Livingston and Absaroka subprovinces to show the relations of these characteristically high-potash rocks. Data for the Livingston subprovince were taken from Merrill's (1895) report on the Bozeman area, those for the Absaroka subprovince from Larsen's (1940) report on the petrographic province of central Montana. The concentration of the analyses of the rocks of the Big Belt Mountains indicates comparatively little differentiation and, probably, eruption within a very short interval of time.

Using the data of the variation diagram (Fig. 4), estimates of the compositions and norms of the materials necessarily added to or subtracted from the Livingston

parent magma to produce its mafic and felsic differentiates show that, for the ultramafic end, calcic plagioclase, diopside, hypersthene, and olivine must be added to the parent melt, but, for the felsic end, these same materials must be subtracted. If material is added to the magma to form the felsic rocks it must be silica-deficient and contain orthoclase, leucite, nepheline, sodium metasilicate, and diopside.

TABLE 5.—*Estimated composition of the material subtracted from the magma of the Big Belt Mountains*
Using both chemical and mineralogical data

	Mineralogical	Chemical
SiO_2	49.1	53.3
Al_2O_3	18.3	18.4
FeO	8.9	11.1
MgO	8.1	4.3
CaO	11.7	10.8
Na_2O	3.2	2.6
K_2O	0.0	0.0
	99.3	100.5

To produce the alkalic lavas of the Big Belts from a parent magma of the San Juan type, labradorite, diopside, hypersthene, olivine, and, for high-soda rocks, small quantities of quartz and corundum must be removed. If material were added to obtain these rocks it would include orthoclase, anorthite, leucite, nepheline, diopside, wollastonite, and sodium metasilicate. Analogous results hold if the mafic rocks of the Big Belts are made into felsic rocks by addition or subtraction of material. In all cases, if subtraction goes on, 50 per cent of the original magma is used up; if addition operates, only 30 per cent, on the average, is added to the initial melt.

Using the optical data on the phenocrysts of the Big Belt rocks to calculate their approximate compositions, and assuming these minerals to have settled in the proportions of 60 per cent labradorite, 20 per cent augite, and 20 per cent olivine, we arrive at the estimated composition of the material subtracted from the magma (Table 5). With it is the calculated composition of material subtracted from a trachybasalt analysis to produce a syenogabbro.

The chief difference in the two analyses is in the relative amounts of MgO and SiO_2, possibly because the estimated composition of the augite is incorrect, inasmuch as Winchell's (1933) data are admittedly insufficient to insure accuracy. The otherwise close correspondence lends strong support to the idea that the phenocrysts actually settled from the parent melt.

Omitting a detailed consideration of the many theories proposed for the origin of alkalic rocks, simple crystal settling alone seems applicable to the Big Belt igneous suite. The evidence favoring this theory rests not only on the congruity of the chemical calculations with the observed mineralogical data but also on the rocks themselves, which exhibit no profound differences and show, as differentiation proceeds, only a gradual and steady decrease in mafic minerals and calcic plagioclase and a corresponding increase of orthoclase.

These results agree with those obtained by Larsen (1940) in his study of the Montana petrographic province. The relation of the alkalic rocks of the province to the orogeny is noteworthy. Those of the Highwood, Absaroka, Little Belt, Castle, Crazy, Sweet Grass, and Yellowstone subprovinces were erupted during periods of very slight tectonic disturbance in the Tertiary. Those of the Big Belt and Livingston areas, both upper Cretaceous, preceded the main diastrophism. The Boulder batholith, midway in age between the two groups, is probably in part contemporaneous with the orogeny. Assimilation, rather than crystal fractionation, seems to have been of major importance there.

TABLE 6.—*Specific gravity, optical, and spectrographic data for the analcimes of the Big Belt, Highwood, and Bearpaw Mountains, Montana*

	Big Belt	Highwood	Bearpaw	Lake Superior
Sp. Gr.	2.286	2.270	2.302	2.240
n	1.493	1.493	1.494	1.486
Wt. % K_2O	large	4.48	large	small
Wt. % CaO	1.19	1.19	1.19	0.40
a_0 (length of unit cell)	13.70Å	13.72Å	13.75Å	

ANALCIME PROBLEM

The origin of analcime phenocrysts up to 5 mm. in diameter in some of the trachybasalts of the Big Belts is not explained by any synthetic silicate study. A common assumption is that analcime is secondary and pseudomorphous after leucite. An investigation of igneous analcimes from the Highwood, Bearpaw, and Big Belt Mountains has been made by the author with Mr. Bernard Fisher of the U. S. Geological Survey, and the following results are largely his.

According to Morey and Ingerson (1937), analcime has been synthesized easily at temperatures from 110°C. to 550°C., and from aqueous compounds of widely differing compositions. Schairer and Bowen (1935), however, reported no $NaAlSi_2O_6$ in their study of the system silica-carnegieite-kaliophilite, so that experimentally, at least, water seems essential to the formation of analcime. Friedel's (1896) dehydration studies showed very plainly that it loses all its zeolitic water below 505°C., but Gruner heated analcime to 700°C. and showed by rotation and powder photographs taken at this temperature that no change had occurred in the unit cell. The zeolitic water, therefore, has no influence in determining the body-centered character of the analcime lattice. In Taylor's (1930) early X-ray study sodium could easily be replaced by silver in this lattice, and, if this were done, the 420 reflection was greatly strengthened.

Abnormally strong 420 reflections in the X-ray powder patterns of the Big Belt, Bearpaw, and Highwood analcimes indicated the probability that metals other than sodium, aluminum, and silicon were abundant. Specific gravities, indices of refraction, and spectrographic analyses (by J. C. Rabbitt of Harvard) corroborated the X-ray data, and the results are shown in Table 6.

Specific gravities were determined pycnometrically. The sample from the Highwood Mountains (F.B.H. 169) has been described by Larsen and Buie (1938), and a chemical analysis has been published. The K_2O quoted for the other analcimes could not be determined accurately because of strong interference by the cyanogen band in the critical portion of the spectroscopic plate. The order of magnitude of the potash content for the Big Belt and Bearpaw analcimes closely approximates, however, that of the analyzed Highwood specimen. The Lake Superior analcime came from an amygdaloidal cavity and has been used for comparison.

The significant feature of the igneous, as contrasted with the amygdaloidal, analcime is its high content of extraneous metals, which makes it probable that the igneous analcimes formed at relatively elevated temperatures.

In all three areas the igneous analcimes investigated show no evidence of origin through replacement of leucite or that the smaller orthorhombic cell of leucite has inverted to the isometric cell of analcime. Leucite is absent from all the Big Belt rocks and very sparingly present in the eastern Bearpaws. The data presented above and these field facts do not demand that the analcime did not crystallize at magmatic temperatures, perhaps even in a very nearly anhydrous state, or that its development as phenocrysts predicates the former presence of leucite.

The high-potash analcimes approach pseudoleucite in composition, and a genetic origin of the latter from the former is not wholly improbable.

SUMMARY

The exposed stratigraphic column at the northern end of the Big Belt Range consists of 1500 feet of Mississippian limestones and shales unconformably overlain by 3550 feet of Mesozoic sandstones, shales, and limestones, and 3200 feet of upper Cretaceous trachybasalts. Part of the Algonkian Belt series crops out on the western edge, but most of the Belt strata and the Cambrian and Devonian formations are unexposed.

The thick volcanic sequence of flows, breccias, and agglomerates is folded along northwesterly trending axes. An imbricate zone on its western margin has formed along the plunging anticlinal structure of the main Big Belt uplift. The Lewis overthrust is at the western boundary of the structurally complex zone.

The potash-rich basalts which form the great volcanic pile of the mountains are petrologically similar to the absarokite-shoshonite-banakite series of the Absaroka Range and to the potash-rich Livingston volcanics (upper Cretaceous) of southwestern Montana. The extrusive rocks are restricted to orthoclase basalts and analcime trachybasalts, but the intrusives grade from gabbros to quartz monzonites. Calculations based on mineralogical and chemical data indicate that crystal settling was the dominant cause of the differentiation within this magmatic series. Petrologic, X-ray, spectroscopic, optical, and gravitative data on the unusual analcime phenocrysts which characterize portions of the volcanics indicate that the analcime crystallized as a primary mineral.

Most of the extrusives are considered to have been erupted from the Three Sisters composite stock in the east-central part of the range. The dike swarm radiating

from this stock is considered to have been emplaced laterally from the stock rather than vertically from beneath because dikes do not intrude the Paleozoic strata. The sills and laccoliths of the mountains and outlying plains have been produced by dikes which bent from a vertical to a horizontal attitude. A peculiar type of layering is described within the Square Butte sill, but its cause is inexplicable.

The groundmass of almost all the igneous rocks is zeolitized. This is possibly due to ground-water alteration of the porous volcanics before cooling, as the zeolitization bears no relation to the intrusive rocks. Oxidation of iron and hydration accompany zeolitization, but no constituent is added to or subtracted from the rocks.

The physiographic features of interest within the northern Big Belts are (1) the production of a water gap through the range by damming and spilling over of the Snake River drainage during the Oligocene, (2) the development of three pediments, Miocene to Pleistocene, and (3) the accumulation of a few hundred feet of sand along the Missouri River valley in the Pleistocene when the valley constituted a part of glacial Lake Great Falls.

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STATE OF NEW HAMPSHIRE, HIGHWAY DEPARTMENT
FREDERIC E. EVERETT, COMMISSIONER AND STATE ENGINEER

(Plymouth)
71° 30'
25'
(Mt Chocorua)
20'
43° 45'
40'
35'
(Holderness)
MOULTONBORO
CENTER HARBOR
MEREDITH
LACONIA
WINNIPESAUKEE
MERRIMACK
BELKNAP
CARROLL CO
Center Harbor
Meredith
Moultonboro
Lees Mills
Lees Pond
Lake Kanasatka
Whitebacks
Pansy I
Hollingsworth Hill
Joes I
Poplar I
Brown Bay
Long Pt
Fivemile I
Sixmile I
Long Island
Little Whortleberry I
Steamboat I
Jolly I
Mark I
Little Camp I
Mink I
Camp I
Forty Is
Round I
Sandy I
Ragged
Little Bear Island
East Bear Island
Dollar I
Bear Island
Gowanors Island
The Witches
Locke I
Rock I
Belknap Pt
Belknap Point
Sanders Bay
Meadow Brook
Locke Hill
Diamond I
Lily Pond
Lily Pond
Meadow River
Marine Bk
Boston
Mountains
Pinnacle Hill
Pinehurst Hill
Oak I
Pinehurst I
Spindle Pt
Eagle Is
The Weirs
Paugus Bay
Twomile I
Threemile I
Finell
Meredith Hill
Holman Hill
Beaver Is
Wallace I
Minister I
Black Cat I
Hall I
Becky's Garden
Lakeview
Wamepesunk
Cooks Pt
Steamboat
Route
wad
mpg
Di
Ch
Ch
Ch

NEW HAMSHIRE
WINNIPESAUKEE QUADRANGLE

CARBONIFEROUS OR TRIASSIC

YOUNGER THAN LOWER DEVONIAN, PROBABLY CARBONIFEROUS

White Mountain Magma Series — age of rocks at Red Hill relative to those of Belknap and Ossipee Mountains is uncertain

Belknap Mountains and Ossipee Mountains

Red Hill

LEGEND

METAMORPHIC AND IGNEOUS ROCKS

Trap
(Fine-grained, dark trap, composed of andesine, pyroxene, and ilmenite.)

Fine-grained quartz syenite
(Fine-grained, gray quartz syenite, composed of microperthite, hastingsite, quartz, and biotite. The hastingsite occurs as needles nearly one-eighth of an inch long.)

Medium-grained syenite
(Medium-grained, gray to pink syenite, composed of microperthite and hastingsite.)

Nepheline-sodalite syenite
(Medium- to coarse-grained, light-gray syenite, composed of the following minerals in varying proportions: microperthite, nepheline, sodalite, and black stubby crystals of hastingsite.)

Coarse-grained syenite
(Coarse-grained, gray to yellowish syenite, composed of microperthite (crystals average one-half inch in length and reach as much as two and one-half inches), hastingsite, and locally small amounts of quartz and biotite.)

Trap-syenite breccia
(At Poorfarm Brook a dike-like mass of fine-grained, black trap with numerous inclusions of syenite.)

Vent-agglomerate
(Fragments of Meredith porphyritic granite, pink quartz syenite, syenite, and Conway granite in a matrix of explosive material, the matrix consisting of microperthite, oligoclase, and quarts.)

Conway granite
(Medium- to coarse-grained, pink biotite granite, composed of microperthite, quartz, oligoclase, and biotite.)

Albany porphyritic quartz syenite
(Medium-grained, pink to gray quartz syenite; phenocrysts chiefly alkali feldspar, three-eighths of an inch long, with some rounded grains of quartz; groundmass composed of alkali feldspar, quartz, and hastingsite. Small dark inclusions common.)

Subporphyritic quartz syenite
(Medium-grained, subporphyritic, pink to gray quartz syenite; phenocrysts chiefly alkali feldspar; groundmass composed of microperthite, hornblende, quartz, and biotite. Dark inclusions common.)

Pink quartz syenite
(Medium-grained, pink quartz syenite, composed of microperthite, hornblende, and quartz.)

Syenite
(Coarse-grained, light-gray syenite; composed of microperthite and hornblende.)

Augite monzodiorite
(Variable but mainly medium-grained, light-gray to gray monzodiorite, composed of oligoclase, microperthite, hornblende, biotite, augite, and locally quartz.)

Monzodiorite
(Medium-grained, gray monzodiorite, composed of

LATE DEVONIAN OR CARBONIFEROUS

Magma series uncertain

Hornblende gabbro
(Coarse-grained, gray to dark-gray gabbro, consisting of variable proportions of labradorite and hornblende. Relationship to other rocks not well revealed.)

YOUNGER THAN LOWER DEVONIAN, PROBABLY LATE DEVONIAN

New Hampshire Magma Series

Concord granite
(Fine- to medium-grained, foliated, light-gray to light-brown granite, composed of orthoclase, quartz, oligoclase, muscovite, and biotite.)

Winnipesaukee quartz diorite
(Medium-grained, slightly foliated, gray quartz diorite, including some granite, granodiorite, and diorite. The principal minerals are oligoclase or andesine, quartz, orthoclase, biotite, and locally hornblende. wm : area in which blocks of Meredith porphyritic granite are included in the fine-grained phase of Winnipesaukee quartz diorite.)

Meredith prophyritic granite
(Coarse-grained light-gray, porphyritic granite, composed of potash feldspar in phenocrysts up to two inches long, quartz, oligoclase, and biotite. Muscovite, garnet, and sillimanite are present at many places. Inclusions of schist common. The Meredith porphyritic granite is essentially the same rock as the Kinsman quartz monzonite; the Meredith is richer in phenocrysts of potash feldspar.)

DEVONIAN ?

Lower ?

Littleton (?) formation
(Fine- to coarse-grained, contorted, gray to rusty-brown metamorphic rocks of the following types: muscovite schist, biotite schist, muscovite-biotite schist, garnet-mica schist, quartz-mica schist, and mica-sillimanite schist.)

METAMORPHIC ZONES
(All the schists show high-grade metamorphism.)

CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

Indefinite as sharp contact is lacking

SPECIAL SYMBOLS
(In the Littleton formation the bedding is generally parallel to the schistosity.)

Strike and dip of bedding, including inverted and normal strata

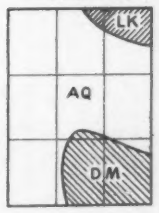
Strike of vertical beds

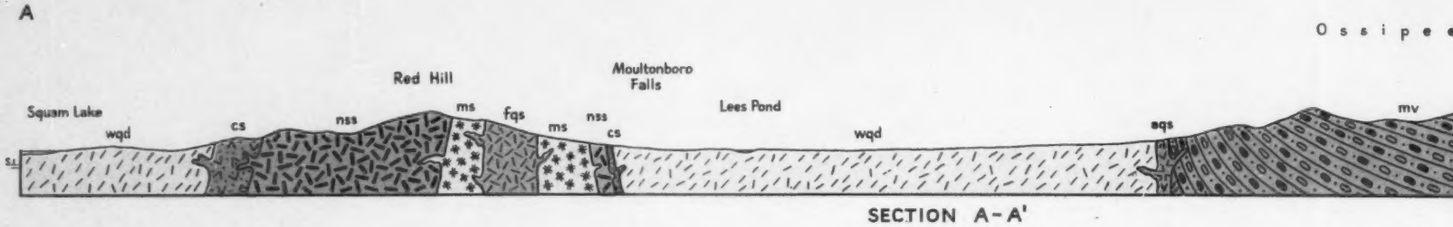
Strike and dip of foliation and schistosity.

Strike of vertical foliation and schistosity

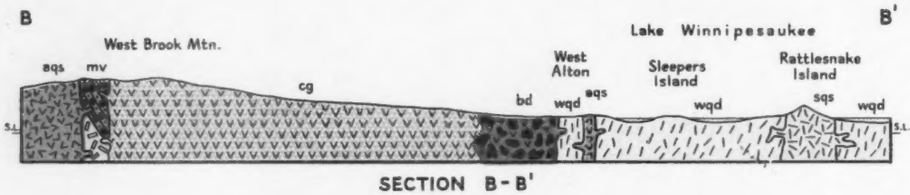
Horizontal foliation

Silicified zones

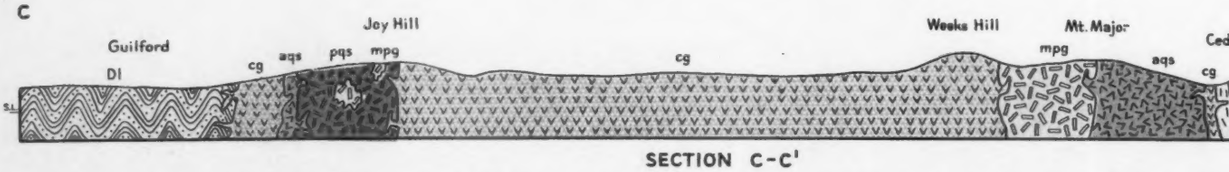




SECTION A-A'



SECTION B-B'



SECTION C-C'

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WINNIPESAUKEE QUADRANG

Topographic base by U. S. Geological Survey, 1906-1907

Scale 1/62500

1 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet.

Datum is mean sea level.

White Mountains

Belknap

am

Augite monzodiorite

(Variable but mainly medium-grained, light-gray to gray monzodiorite, composed of oligoclase, microperthite, hornblende, biotite, augite, and locally quartz.)

m

Monzodiorite

(Medium-grained, gray monzodiorite, composed of andesine, microperthite, hornblende, biotite, and quartz.)

bd

Breccia of diorite in Conway granite

(Fine- to coarse-grained, gray to dark-gray diorite, composed of oligoclase, hornblende, biotite, and pyroxene. Present only as inclusions, a few inches to several hundred feet long, in a matrix of syenite or Conway granite.)

g

Gabbro

(Coarse-grained, blotchy, dark gabbro, composed of labradorite, hornblende, and pyroxene.)

mv

Moat volcanics

(Interbedded flows, tuffs, and breccias of rhyolitic, andesitic, and basaltic composition.)

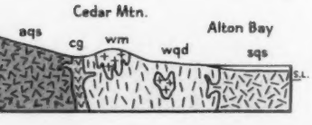
D M

A'

s i p e e M o u n t a i n s

B'

C'



RANGLE, NEW HAMPSHIRE

Geology by Alonzo Quinn, David Modell, and Louise Kingsley. Directed by Marland Billings. Geology surveyed 1929-1938, with the aid of grants from the Holden Fund and the Associates in Science of Harvard University. Published 1940.

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MAGMATIC CONTRASTS IN THE WINNIPESAUKEE REGION, NEW HAMPSHIRE

BY ALONZO QUINN

CONTENTS

Page
Abstract 474
Introduction 474
Acknowledgments 474
Littleton(?) formation 475
New Hampshire magma series 476
General statement 476
Meredith porphyritic granite 476
General statement 476
Age relations 476
Megascopic and microscopic characters 477
Origin 478
Winnipesaukee quartz diorite 478
Concord granite 479
Hornblende gabbro 479
General characteristics of the New Hampshire magma series 479
Origin of the New Hampshire magma series 479
General considerations 479
Origin by differentiation of basaltic magma 482
Origin by melting of sial 482
Origin by assimilation 484
Conclusion 484
White Mountain magma series 484
General statement 484
Ossipee Mountains 484
Belknap Mountains 485
Red Hill 485
Dikes 485
Origin of the White Mountain magma series 485
General considerations 485
Change by subtraction 488
Change by addition 492
Conclusion 492
Summary and conclusions 494
References cited 495

ILLUSTRATIONS

Figure Page
1. Index map 475
2. Variation diagram of White Mountain magma series 482
3. Variation diagram of New Hampshire magma series 483
Plate Facing page
1. Geologic map and structure sections of Winnipesaukee quadrangle 473

ABSTRACT

In the Winnipesaukee area of central New Hampshire two groups of igneous rocks are strikingly different. The New Hampshire magma series is probably late Devonian, and the White Mountain magma series is probably Mississippian. The contrasting characteristics of these two magma series are tabulated.

The New Hampshire magma series here consists of large partially concordant intrusions of Meredith porphyritic granite and Winnipesaukee quartz diorite and small intrusions of hornblende gabbro and Concord granite. Conclusive evidence of the origin of this series is lacking, but some evidence indicates that it has been derived by melting of sial.

The White Mountain magma series includes a great variety of extrusive and intrusive rocks in the Ossipee Mountains, the Belknap Mountains, and Red Hill. Cauldron subsidence is an important method of intrusion. The order of intrusion, from oldest to youngest, is generally gabbro, diorite, monzodiorite, syenite, quartz syenite, granite. Calculations indicate the nature of the materials which could be subtracted or added, to cause the changes shown by this sequence. These calculations suggest that neither subtraction, in the form of crystal fractionation, nor addition, in the form of assimilation or pure melting, is the sole explanation, but that both played a part. Crystal fractionation probably dominated up to the stage of syenite, and assimilation probably assumed increased importance in the change from syenite to granite.

INTRODUCTION

In the vicinity of Lake Winnipesaukee in central New Hampshire (Fig. 1) are two contrasting groups of igneous rocks. On the basis of work by Billings (1935) these groups of rocks have been termed the New Hampshire magma series and the White Mountain magma series. The New Hampshire magma series was intruded in the later stages of Devonian orogeny and is probably late Devonian. The White Mountain magma series formed after erosion had exposed plutonic rocks of the New Hampshire series and is, therefore, considerably younger. Williams and Billings (1938, p. 1025) suggest a Mississippian age for it. Both series intrude schist of the early Devonian Littleton formation.

In the following pages the schist is described briefly, and the rocks of the New Hampshire magma series in more detail. The rocks of the White Mountain series in the Winnipesaukee quadrangle have been described in some detail (Kingsley, 1931; Modell, 1936; Quinn, 1937) and are reviewed only in a general way. The origin of the two magma series is discussed.

ACKNOWLEDGMENTS

During a number of years of work in and near the Winnipesaukee area, the writer received help from several people and organizations. The colored geologic map of the Winnipesaukee quadrangle (Pl. 1) was furnished by the State of New Hampshire Highway Department under the direction of Mr. Frederic E. Everett. Sincere gratitude is due Professor E. S. Larsen and Professor Marland Billings of Harvard University for discussion of the problems, although they may not agree with all the conclusions. Professor Billings has given generously of his time in critical reading of the manuscript. The Associates in Science of Harvard University provided funds for some of the field work. Thanks are due Professor J. W. Goldthwait of Dartmouth College for numerous courtesies during the field work. Mr. R. N. Palmer, who has a summer home at Keewayden in Wolfeboro, furnished boat transportation to several islands in Lake Winnipesaukee. The analyses of rocks at Red Hill (Quinn, 1937) were paid for by a grant from the Penrose Bequest of The Geological Society of America. Field work in the eastern part of the area, more especially in the Wolfeboro quadrangle to the east, was supported by a grant from the Bache Fund of the National Academy of Science.

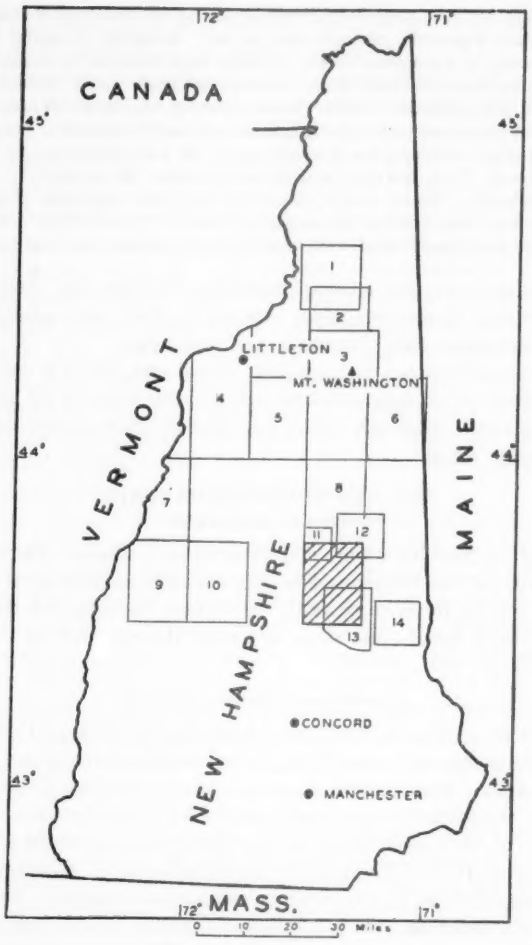

FIGURE 1.—*Index map*

Showing location of Winnipesaukee quadrangle and other areas in New Hampshire which have been described recently.

Winnipesaukee area shaded; 1. Percy area (R. W. Chapman, 1935); 2. Pliny area (R. W. Chapman, 1942); 3. Mt. Washington area (Billings, 1941); 4. Littleton-Moosilauke quadrangle (Billings, 1937); 5. Franconia quadrangle (Williams and Billings, 1938); 6. North Conway quadrangle (Billings, 1928); 7. Mt. Cube area (Hadley, 1942); 8. Mt. Chocorua quadrangle (Smith, Kingsley, and Quinn, 1939); 9. Mascona quadrangle (C. A. Chapman, 1939); 10. Cardigan quadrangle (Fowler-Lunn and Kingsley, 1937); 11. Red Hill (Quinn, 1937); 12. Ossipee Mountains (Kingsley, 1931); 13. Belknap Mountains (Modell, 1936); 14. Merrymeeting Lake area (Quinn and Stewart, 1941).

LITTLETON(?) FORMATION

West of the Belknap Mountains and on Wolfeboro Neck are areas underlain by schist and quartzite of the Littleton(?) formation, the oldest rocks of the Winnipesaukee quadrangle.

The schist is gray, but most outcrops are brown because of the weathering of small grains of sulphides. The texture is generally medium-coarse to fine. Schistosity is parallel to the original bedding, which commonly is complexly folded. Different small beds may be classified as muscovite schist, biotite schist, muscovite-biotite schist, muscovite-sillimanite schist, sillimanite-mica schist, garnet-mica schist, and quartz-mica schist. Quartz, muscovite, and biotite are present in almost all types, and sillimanite is common, although in many places it has been altered to muscovite. Garnet is common in crystals of millimeter size or a little larger. In a few places garnets 3 millimeters or more across are found. Pyrite and other sulphides are abundant but are generally not megascopically visible. Andalusite, chlorite, sericite, corundum, pyrrhotite, orthoclase, plagioclase, zircon, sphene, and spinel have been identified microscopically (Modell, 1936, p. 1890). Almost everywhere the schist is cut by small granite dikes, small irregular pegmatite dikes, and quartz veins.

The texture, structure, and mineral composition indicate that these rocks are in the high-grade zone of metamorphism. Originally, they were sedimentary rocks, mainly shale, sandstone, shaly sandstone, and sandy shale.

The schist cannot be dated with certainty in this area, but it is correlated tentatively, on the basis of lithology and structure, with the schist of the early Devonian Littleton formation. Hitchcock (1877) and Modell (1936) called these rocks the Rockingham mica schist.

NEW HAMPSHIRE MAGMA SERIES

GENERAL STATEMENT

The rocks of the New Hampshire magma series are plutonic. The Meredith porphyritic granite and the Winnipesaukee quartz diorite underlie great areas (Pl. 1). The Concord granite forms three small masses near Laconia, and there is a small body of hornblende gabbro just south of Center Harbor. All are considered late Devonian (Billings, 1935, p. 26).

MEREDITH PORPHYRITIC GRANITE

General statement.—The Meredith porphyritic granite (Billings, 1928, p. 83) crops out mainly along the west edge of the Winnipesaukee sheet from Center Harbor almost to Laconia. It is exposed in many road cuts between Center Harbor and The Weirs; in the Meredith trap quarry, near Little Pond between Center Harbor and Meredith; on the tops of several of the higher hills, such as Sunset Hill, Gilman Hill, and Pinnacle Hill. This granite is correlated with the Kinsman quartz monzonite (Billings, 1935, p. 27, 28) which is widespread throughout New Hampshire. Daly (1897) concluded that the parallelism of feldspar crystals indicates magmatic flowage rather than metamorphism, a conclusion supported by the present investigation.

Age relations.—The Meredith porphyritic granite is clearly younger than the Littleton(?) schist, as it cuts and includes pieces of the schist. Excellent examples of schist inclusions may be seen at the top of Pinnacle Hill. In the main, the porphyritic granite cuts across the foliation, and the phenocrysts lie parallel to the contract rather than to the schistosity.

The age relations of the Meredith porphyritic granite to other rocks of the New Hampshire magma series are not shown in such a way as to give a completely satisfactory conclusion. Daly (1897, p. 707–710) described a number of outcrops, several within a few miles of The Weirs, which convinced him that the Meredith porphyritic

granite is younger than the Winnipesaukee quartz diorite. It appears to the writer that there is better evidence to the contrary, however.

Among the features described by Daly are masses of rock, which he referred to both as "Lake Winnipiseogee gneiss" and as schist, cut by apophyses of the porphyritic granite. The localities described are on Governors Island, Spindle Point, and an inlet which is probably the same as Maiden Lady Cove of the present map. Between Maiden Lady Cove and The Weirs are good cuts on a new road which runs just above the railroad. In these, masses of schist are included in the Meredith porphyritic granite. At places the schist is well foliated and undoubtedly of the Littleton(?) formation, while at others the rock is so massive as to resemble the Winnipesaukee quartz diorite. The writer believes that it is this rock rather than the Winnipesaukee quartz diorite which is cut by apophyses of the Meredith porphyritic granite. Furthermore, on the southwest shore of Sphindle Point is what appears to be a small dike of Winnipesaukee quartz diorite cutting the Meredith porphyritic granite at an acute angle to the flowage structure of the porphyritic granite. The gradational contact at Center Harbor (Daly, 1897, p. 710) agrees with either age relationship. The "horse" on Governors Island may be interpreted as an intrusion of Winnipesaukee quartz diorite.

An additional locality of significance is Cedar Mountain near the southeast corner of the Winnipesaukee map. On the upper part of this hill is an intrusion breccia of the Meredith porphyritic granite cut by what appears to be the Winnipesaukee quartz diorite. There can be no doubt that the Meredith porphyritic granite is the older rock, and the younger rock strongly resembles the Winnipesaukee quartz diorite. There is a possibility, however, that the younger rock is a phase of the Concord granite or some other rock. The uncertainties have to do with correlation.

A further evidence of the greater age of the Meredith porphyritic granite is the absence of inclusions of undoubted Winnipesaukee quartz diorite in the porphyritic granite. This negative evidence seems especially strong in view of the fact that inclusions of Littleton(?) schist in the porphyritic granite are numerous.

Megascopic and microscopic characters.—The Meredith porphyritic granite is white to light gray and is most striking because of the very large crystals of potash feldspar. In groundmass, representing fully half the rock, most of the grains do not exceed 3 millimeters in diameter, but the big feldspar crystals are 2 to 5 centimeters long and 1 to 2 centimeters across. In many places the feldspar phenocrysts are parallel to the walls of the intrusion or to the edges of inclusions of schist, but elsewhere they seem to have random orientation.

The large phenocrysts are Carlsbad twins of potash feldspar. The matrix consists of plagioclase, quartz, biotite, and muscovite. Small garnets are common, and in several places there are small radial blades of sillimanite. At the Meredith trap quarry serpentine may be seen along shear zones.

In thin section, the phenocrysts are seen to be chiefly microcline, but some are orthoclase and some microperthite. Large grains of oligoclase formed early in the solidification of the rock. Myrmekite is late and replaces the edges of the phenocrysts. The oligoclase of the myrmekite is usually surrounded by a thin shell of albite, especially where the myrmekite has replaced potash feldspar. A few grains of biotite are included in the phenocrysts, but much of it is interstitial or has partially replaced the phenocrysts and seems to be late magmatic or hydrothermal. Quartz, muscovite, and sillimanite are late. In summary, accessory minerals, a few oligoclase crystals, and part of the biotite crystallized early; big potash feldspar phenocrysts then solidified; myrmekite, quartz, part of the

biotite, muscovite, and sillimanite formed later as the final magmatic stage or perhaps in part from a still later hydrothermal phase.

The replacement of microcline by myrmekite must have released considerable potash and thus was an additional source for the potash enrichment of schist (Billings, 1938).

The coarse texture of this rock and its variability prevent accurate determination of the mode. A flat surface at the Meredith trap quarry was measured to get the ratio of matrix to phenocrysts, and then sections of the matrix were studied. In the quarry, phenocrysts comprise 40 to 50 per cent of the rock (Table 1). At other places in this quadrangle the phenocrysts are less abundant.

TABLE 1.—*Mode of Meredith porphyritic granite at Meredith trap quarry*

	Per cent by volume
Potash feldspar phenocrysts	45
Plagioclase (oligoclase)	4
Myrmekite	15
Quartz	19
Biotite	9
Muscovite, sillimanite, garnet, accessories	8
	100

Origin.—Daly (1897) demonstrated that the Meredith porphyritic granite is of eruptive origin, and the present investigation agrees. Among the strongest evidences of the igneous origin are the cross-cutting relationships of the porphyritic granite to the older schist, the sharp contacts of schist inclusions, the flowage structure shown by the big phenocrysts, and the great size of the porphyritic granite bodies here and elsewhere in New Hampshire.

Perhaps the main objection which has been raised is that, under a simple igneous origin, the phenocrysts would be among the first crystals formed, whereas potash feldspar is usually late in the crystallization of a granitic rock. Textural evidence described above shows that only a small amount of biotite, oligoclase, and accessories preceded the potash feldspar and that the phenocrysts were rather early in the crystallization. The textural evidence, therefore, does not oppose an igneous origin.

Another evidence which may be used against the igneous origin is the presence of contact zones between the Winnipesaukee quartz diorite and Littleton(?) schist in which are metacrysts of feldspar. Such zones look very much like the Meredith porphyritic granite, but they are of an entirely different order of size, being only a few feet across. Such a small contact zone may be seen near the east edge of the map, at latitude 43° 40′.

The evidence indicates emplacement as a partially liquid mass, in which floated the big feldspar crystals and many inclusions of schist. The solid portion probably did not exceed 40 per cent of the whole. The phenocrysts became oriented by flowage.

WINNIPESAUKEE QUARTZ DIORITE

The Winnipesaukee quartz diorite, the most extensive rock in the area, underlies most of the lake, and its relatively rapid erosion accounts for the lowland. It is the "Winnipiseogee gneiss" of Hitchcock (1877, p. 595–599) and the Chatham of Kingsley (1931) and Modell (1936).

The Winnipesaukee quartz diorite is light gray to dark gray. Grains, commonly 2 to 4 millimeters across, range up to 6 millimeters. Scattered porphyritic phases contain phenocrysts up to 10 millimeters in diameter. This rock is massive or foliated. The chief minerals are plagioclase, orthoclase, quartz, biotite, and muscovite. In most places it is quartz diorite, but locally it is granodiorite, granite, or diorite. Modes and compositions of the plagioclase are given in Table 2. Quartz is interstitial and commonly has an unusually wavy extinction, apparently because of crushing. Biotite usually has β and γ about 1.645, but the range is from 1.640 to 1.655; $2V(-)$ is very small; and the pleochroism is from brown to very pale brown. Biotite in the White Mountain magma series usually has higher indices (Chapman and Williams, 1935, Table 8; Quinn, 1937, p. 390; Quinn and Stewart, 1941). On Mark Island and the southeast part of Bear Island the dark mineral is mainly hornblende with the following optical properties: $\alpha = 1.668$, $\beta = 1.681$, $\gamma = 1.693$, $r > v$, $2V(-)$ moderate to large; X light brownish yellow, Y green, slightly brownish, Z bluish green, $Y > Z > X$; $Y = b$. Accessory minerals are apatite, sphene, zircon, magnetite, ilmenite, garnet, and epidote. This rock generally contains numerous small irregular intrusions of pegmatite and aplite. Most of the pegmatites contain only feldspars, quartz, muscovite, and biotite, but a very few near the end of Long Island and on near-by islands in Lake Winnipesaukee contain greenish beryl and black tourmaline.

The Winnipesaukee quartz diorite occurs as intrusions into the schist and contains inclusions of schist. Locally, it cuts across the foliation of the schist but it tends to be concordant in a large way. Concordance is better shown on the Chocorua quadrangle to the north (Smith, Kingsley, and Quinn, 1939).

CONCORD GRANITE

Three patches of muscovite-biotite granite cut the schist near Laconia. Their age is not known, but they are correlated with the Concord granite of the New Hampshire magma series.

These rather well-foliated intrusions are light gray to light brown. Their texture is fine to medium. The main minerals are orthoclase and microcline, oligoclase, quartz, muscovite, and biotite. Accessory minerals are apatite, zircon, and magnetite.

HORNBLENDE GABBRO

About half a mile south of Center Harbor is a small patch of hornblende gabbro, so poorly exposed that its relations to the surrounding rocks are not known. It appears to be younger than the Meredith porphyritic granite and probably belongs to the New Hampshire magma series.

The rock is gray to dark gray, depending upon whether labradorite or hornblende is dominant. Other minerals, in small amounts, are sphene, chlorite, iron ore, and pyrite.

GENERAL CHARACTERISTICS OF THE NEW HAMPSHIRE MAGMA SERIES

Certain general characteristics of the New Hampshire magma series and particularly the contrasts with the White Mountain magma series have been arranged in Table 3. Published work on other areas indicates that these characteristics apply throughout New Hampshire, as well as to the Lake Winnipesaukee area.

ORIGIN OF THE NEW HAMPSHIRE MAGMA SERIES

General considerations.—The rocks of the New Hampshire magma series form large partially concordant intrusions which were emplaced during the last stages of orogeny, as are many large granitic intrusions.

TABLE 2.—*Modes of the Winnipesaukee quartz diorite*

In per cent by volume

	W1	W2	W11	W17	W19	W27	CM	MI	Modell (1936)				Red Hill extremes
									1	2	3	4	
Potash feldspar	7	37	50	5		3	10	3		15	40	33	
Plagioclase	28	32	23	59	42	61	64	61	70	50	30	34	57–75
	An_{37}	An_{34}	An_{27}	An_{44}	An_{38}	An_{34}	An_{28}	An_{48}	An_{35}	An_{30}			An_{28-40}
Quartz	56	22	17	24	50	29	13	15	15	25	27	30	15–23
Biotite	9	9	10	12	8	7	11	17	15	10	3	3	10–20
Hornblende								2					
Muscovite							2						
Accessories								2					
	100	100	100	100	100	100	100	100	100	100	100	100	

W1 Along road at west shore Mirror Lake.
W2 On Lake Shore road just north of Woodmans Cove.
W11 Road cut near "G" of LONG ISLAND.
W17 On road a quarter of a mile southeast of Lakeview, Moultonboro Neck.
W19 Moultonboro Neck, near southeast contact of body of Meredith porphyritic granite.
W27 Road cut near 564-foot corner south of Red Hill.
Cm Cedar Mountain.
MI Mark Island.
Modell. *See* Modell (1936, p. 1893).
1 Minge Cove.
2 South shore Rattlesnake Island.
3 Lake Shore road opposite Minge Cove; plagioclase is oligoclase.
4 Injection in schist; location not given; plagioclase is andesine.
Red Hill. *See* Quinn (1937, p. 378).

The compositions and variations are indicated in Figure 3. In making this diagram, analyses from various parts of New Hampshire were used because only two from the Winnipesaukee area are available. Such use of analyses from widely separated localities might be expected to cause irregularity of the curves, for, as illustrated

TABLE 3.—*Magmatic contrasts*

New Hampshire magma series	White Mountain magma series
No rocks of felsitic texture.	Felsitic texture and felsitic-porphyritic texture common.
Commonly foliated and probably came in under pressure, associated with orogeny.	Commonly massive; and miarolitic cavities common. No compression indicated.
Large bodies, partly concordant intrusions, no extrusions.	Small bodies, discordant intrusions, ring dikes, and stocks; extrusions commonly associated; cauldron subsidence an important mechanism of intrusions.
"Normal" content of alkalies.	High content of alkalies, especially soda.
Most rocks have quartz.	Some silica-poor rocks, such as syenite and feldspathoid syenite.
Muscovite common.	Muscovite usually absent.
Pyroxene and amphibole usually absent.	Pyroxene and amphibole common.
Olivine, fayalite, hastingsite, nepheline, riebeckite, hedenbergite, analcime, sodalite absent.	Olivine, fayalite, hastingsite, nepheline, riebeckite, hedenbergite, analcime, sodalite present.
Much pegmatite of simple mineral composition.	Little pegmatite.
No inclusions of dark differentiates; many schist inclusions in Meredith porphyritic granite.	Many inclusions of dark differentiates.
Relatively small quantities of gabbro and other mafic rocks.	Basaltic extrusions abundant; gabbroic intrusions present, but not abundant.
Little variation, and that within a restricted range.	Much variation, from gabbro to granite.
Variations and order of intrusion not systematic.	Order of intrusions systematic, from early gabbro to late granite; order of extrusions not systematic.
Few dikes (other than aplite and pegmatite).	Dikes numerous and of many types of rock.

by the White Mountain magma series (Fig. 2), the curves for analyses from a restricted area are more regular. The curves in Figure 3 are remarkably regular, however, except for BM4, and the location and geologic relationships of this specimen are not well known (Pirsson and Washington, 1906, p. 466). It appears to be from a small mass, perhaps an aplite dike, intrusive into schist. Its lack of agreement with the other rocks is, therefore, not surprising.

These curves suggest that this series is of rather restricted composition, for the curves do not extend into the left part of the diagram. The curves could be extended to the left by analyses of known rocks of this series, but mafic rocks are very uncommon in this series. In the Winnipesaukee quadrangle approximately 5000 times as much area is underlain by granitic rocks as by hornblende gabbro.

Three theories are postulated for the origin of the large granitic bodies commonly associated with the late stages of orogeny; (1) They are the upper lighter fractions of great masses of differentiated basaltic magma; (2) they are remelted sial; (3) differ-

entiating basaltic magma assimilated sialic material, thereby increasing the amount of the light fraction.

Origin by differentiation of basaltic magma.—Opposing the origin by differentiation of basaltic magma are the scarcity of basalt or gabbro, the abundance of granitic

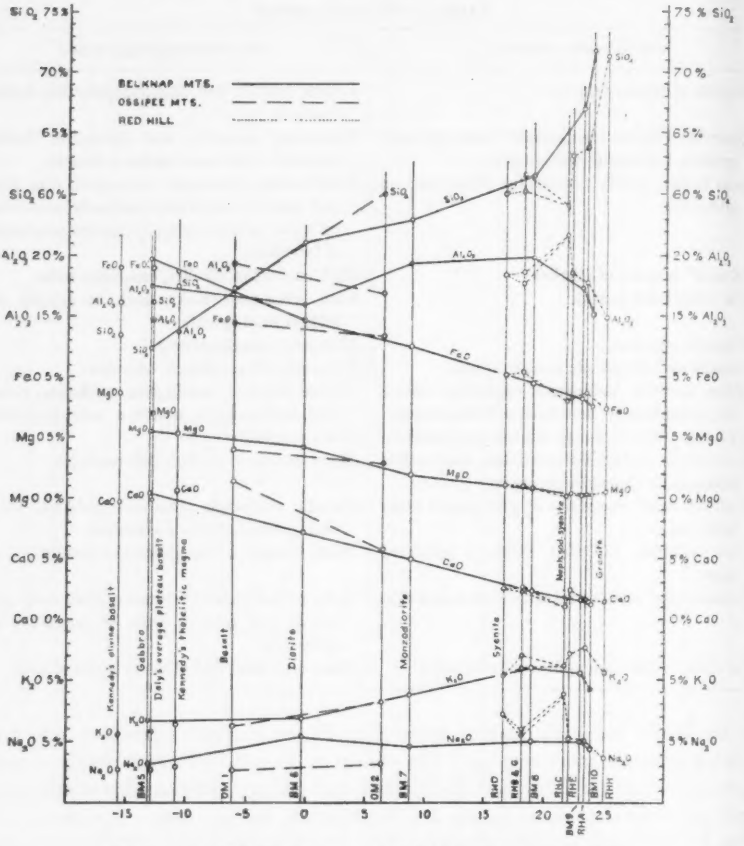

FIGURE 2.—*Variation diagram of White Mountain magma series*
Abscissa ½ SiO_2 + K_2O − FeO − MgO − CaO (Larsen type diagram)

bodies, the restricted composition, and the lack of system in the order of intrusion. The present erosion surface may give a wrong idea of the relative amounts of different rocks, however. Furthermore, the regularity of the lines in Figure 3 suggests igneous rocks derived by differentiation.

Origin by melting of sial.—Daly (1933, p. 311–313) argues that the melting of sial during orogeny best explains many plutonic rocks. Eskola (1932) also argues for the origin of granite magma in part by refusion of the sial, although also in part by

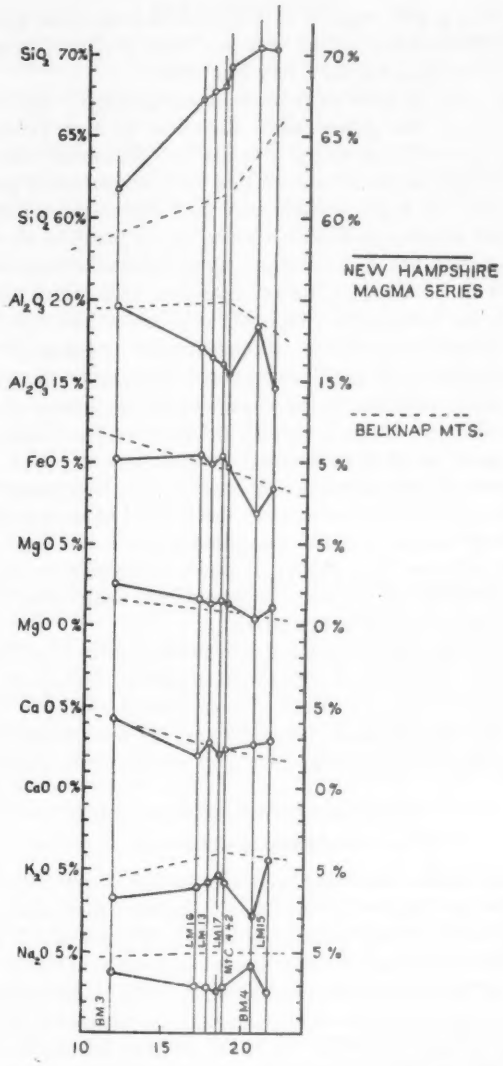

FIGURE 3.—*Variation diagram of New Hampshire magma series*
Abscissa ½ $SiO_2 + K_2O - FeO - MgO - CaO$. (Larsen type diagram)

"sweating out" of the sima. He suggests that some magma series may have a great deal of melted sima and little sial, and some may have much sial and little sima. The New Hampshire magma series corresponds to his type with much sial and little

sima. Lane (1931, p. 820) suggests that crystals as large as the phenocrysts of the Meredith porphyritic granite formed because a lower portion of the sial remained at its crystallization temperature for a very long time.

Opposing the origin of these rocks by the melting of sial is the presence of beryl and tourmaline in a few pegmatities. Elsewhere in New Hampshire, complex pegmatites are apparently associated with the New Hampshire magma series, as at Grafton Center where pegmatite is associated with the Bethlehem gneiss (Chapman, 1939, p. 147–148). It seems unlikely that such concentrations of beryllium and boron, or the rare elements at Grafton Center, can originate by the melting of sediments of a geosyncline, but these elements may be common in the lower sial, and their presence cannot be considered proof of connection with primary basaltic magma. Furthermore, in the fusion of sial, very probably connections with deeper layers of primary basalt would be established. Such is implied by Eskola (1932).

Origin by assimilation.—In the Meredith porphyritic granite, garnet and sillimanite and the many schist inclusions indicate considerable contamination and suggest an origin by assimilation. There is no such evidence for the Winnipesaukee quartz diorite or the small bodies of gabbro and Concord granite. There would be little difference between magmas produced by melting and those resulting in part from assimilation, except that with assimilation there would be more evidence of connection with primary basaltic magma—associated gabbroic intrusions, basaltic dikes, and systematic variation from gabbro to granite. Assimilation might be great or small, and the resulting series of rocks might resemble those caused by melting or by differentiating basaltic magma, respectively.

Conclusion.—No definite conclusion about the origin of the New Hampshire magma series is drawn, for it was probably produced at depths. To a large extent the solution of this problem must be obtained from general evidence. It seems worth pointing out, however, that these rocks show very few features which indicate a connection with primary basalt and that origin by melting of sial seems likely.

WHITE MOUNTAIN MAGMA SERIES

GENERAL STATEMENT

Three areas of White Mountain magma series rise above the lowlands of Lake Winnipesaukee—the Ossipee Mountains, the Belknap Mountains, and Red Hill, all only partly in the Winnipesaukee quadrangle. Pirsson and Washington (1905; 1906; 1907) published enough descriptions and analyses to indicate the nature of the magma series at the Belknap Mountains and Red Hill. Later papers give further descriptions and analyses of the rocks, more detailed maps, and structural studies of the Ossipee Mountains (Kingsley, 1931), the Belknap Mountains (Modell, 1936), and Red Hill (Quinn, 1937). These areas will, therefore, be only reviewed.

OSSIPEE MOUNTAINS

At the Ossipee Mountains there is a great mass of Moat volcanics, a ring dike of Albany quartz syenite, and an interior stock of Conway granite. This area is notable for the unusually complete ring dike, the first in this country to be described.

BELKNAP MOUNTAINS

The Belknap Mountains are somewhat more complex, both in the variety of rock types and structurally. Extrusions are indicated by one area of Moat volcanics south of Round Pond near the south edge of the map and an area of vent agglomerate about 1½ miles northeast of Gilford. Among the intrusive rocks are gabbro, diorite, monzodiorite, augite monzodiorite, syenite, three types of quartz syenite, and Conway granite. In the north and northwest part of the area is a remarkable mass of intrusion breccia with diorite fragments in Conway granite and, on Poorfarm Brook, a smaller area of intrusion breccia with syenite fragments in trap matrix. Both may be seen near the ski jump near Poorfarm Brook.

The intrusions are ring dikes and stocks. The inner margin of the ring dike of Albany quartz syenite is marked by a zone of mylonite.

RED HILL

At Red Hill six types of plutonic rocks have been described: coarse-grained syenite, nepheline-sodalite syenite, medium-grained syenite, fine-grained quartz syenite, medium-grained quartz syenite, and fine-grained granite. The last two are present only in the Mt. Chocorua quadrangle to the north. This locality is widely known for the nepheline-sodalite syenite. The intrusions are concentric or round. No extrusions are known here.

DIKES

Associated with the larger intrusions of the White Mountain magma series are numerous dikes, apparently most abundant at Red Hill and at the Belknap Mountains, but common throughout the area. Mafic dikes are most numerous. The following dikes have been described: aplite, quartz bostonite, camptonite, augite camptonite, augite-olivine camptonite, diorite, granite porphyry, kersantite, paisanite, spessartite, syenite porphyry, quartz syenite porphyry, tinguaite, and vogesite. The Meredith trap quarry, near Little Pond about half way between Meredith and Center Harbor, is in an unusually irregular flat-lying dike. Quarrying has exposed the dike and a great deal of the Meredith porphyritic granite.

ORIGIN OF THE WHITE MOUNTAIN MAGMA SERIES

General considerations.—Chapman and Williams (1935) described the chemical and mineralogical characteristics of the White Mountain magma series and discussed a number of the problems of origin. Several analyses published since 1935 give additional information. Analyses of the rocks from the Ossipee Mountains, the Belknap Mountains, and Red Hill are plotted according to the Larsen (1938) method (Fig. 2).

The two analyses for the Ossipee Mountains have little significance. It will be noticed, however, that, although the lines (long dashes) run in about the same directions as the lines for the Belknap analyses (solid lines), several of the points of the two sets of curves differ considerably.

In general, the Belknap curves run from gabbro, through diorite, monzodiorite, syenite, quartz syenite, and, finally, with little curvature, toward granite. This is the order of age, from oldest to youngest, as determined by Modell (1936). An

TABLE 4.—*Analyses used in variation diagrams and in calculations*

	Red Hill (Quinn, 1937)								Belknap Mountains (Modell, 1936)								Ossipee Mts. (Kingsley, 1931)		Littleton-Moosilauke Area (Billings, 1937)				Mt. Cube Area (Hadley, 1942)
	RHA	RHB	RHC	RHD	RHE	RHF	RHG	RHH	BM3	BM4	BM5	BM6	BM7	BM8	BM9	BM10	OM1	OM2	LM13	LM15	LM16	LM17	MtC 442
SiO_2	63.42	59.10	58.30	59.01	62.20	63.84	60.90	70.89	60.68	69.76	43.94	52.95	56.66	60.75	66.14	70.45	49.04	57.85	67.04	69.12	65.52	66.74	67.83
TiO_2	.18	.89	.10	.81	.34	.31	.47	.38	.64	.36	4.13	3.90	1.55	.63	.51	.69	3.74	2.34	.57	.28	.70	.70	.71
Al_2O_3	17.68	17.42	21.38	18.18	18.34	17.54	18.52	14.84	19.22	18.22	16.17	14.96	19.03	19.68	17.04	14.87	18.10	16.30	16.30	14.30	16.72	15.70	15.24
Fe_2O_3	1.80	2.50	1.05	1.63	1.01	1.88	1.41	.24	.78	.25	3.96	2.44	2.67	1.54	1.24	1.07	2.31	3.16	.74	1.32	.60	.85	.55
FeO	2.04	2.96	2.04	3.65	2.26	2.11	3.80	2.12	4.34	1.59	10.06	7.03	4.88	2.98	2.10	1.54	6.69	5.10	3.97	1.91	4.75	4.30	4.04
MnO	.05	.06	tr	.03	.03	.03	.09	.06	.02								.08	.10	.04	.15	.06	.06	.08
MgO	.30	.99	.22	1.05	.42	.47	.92	.54	2.45	.40	5.05	3.86	1.86	.81	.21	.35	3.68	2.68	1.31	1.04	1.55	1.50	1.37
CaO	1.60	2.48	.95	2.40	2.32	.96	2.14	1.36	4.14	2.68	9.59	6.76	4.82	2.29	1.62	1.28	10.72	5.50	2.68	2.68	1.94	1.96	2.26
Na_2O	5.06	5.81	8.66	7.03	5.12	5.41	5.35	3.70	3.56	4.06	2.93	4.95	4.52	4.89	4.88	4.28	2.53	3.00	2.83	2.49	2.76	2.60	2.67
K_2O	7.64	6.92	6.06	5.34	7.01	7.07	5.75	5.50	3.13	2.06	1.51	1.64	3.62	5.90	5.54	4.16	1.09	2.92	4.03	5.44	3.68	4.48	3.96
H_2O+	.46	.52	.45	.50	.49	.43	.71	.42	.91	.50	1.42	.55	.60	.08	.41	.65	1.72	.61	.65	.53	1.21	.79	.81
H_2O-	.03	.04	.35	.15	.07	.10	.04	.03		.15	.13	.09		.24		.50	.12	.10					.02
CO_2		.03									.09								.00	.31	.00	.00	.06
P_2O_5		.09	.04	tr	.11		.02	.04	.04		.69	.76			.08	.04	.08	.13	.04	.12	.23	.11	.25
S	.01	.02			.04		.03	.04											.11	.06	.19	.10	n.d.
SO_3			.08						.03				.07										
BaO		.06	none	.08	.06		.09	.05											.00	.18	n.d.	n.d.	n.d.
ZrO_2		.09	.02	tr	.14			.05				.02							n.d.	.00	n.d.	n.d.	n.d.
SrO				tr																			
Cl		.12	.35	.12	.06																		
	100.27	100.10	100.05	99.98	100.02	100.15	100.24	100.26	99.94	100.03	99.67	99.91	100.28	99.79	99.77	99.88	99.90	99.79	100.31	99.93	99.91	99.89	99.85

WHITE MOUNTAIN MAGMA SERIES
Red Hill (Quinn, 1937)
RHA—Outer coarse syenite. F. A. Gonyer, analyst.
RHB—Nepheline sodalite syenite, foidomorphic gneiss phase. F. A. Gonyer, analyst.

	100.27	100.10	100.05	99.98	100.02	100.15	100.24	100.26	99.94	100.03	99.67	99.91	100.28	99.79	99.77	99.88	99.90	99.79	100.31	99.93	99.91	99.89	99.85

WHITE MOUNTAIN MAGMA SERIES

Red Hill (Quinn, 1937)

RHA—Outer coarse syenite. F. A. Gonyer, analyst.

RHB—Nepheline-sodalite syenite, feldspathoid-poor phase. F. A. Gonyer, analyst.

RHC—Nepheline-sodalite syenite, "Horne quarry". H. S. Washington, analyst.

RHD—Alkalic syenite of Bayley. W. F. Hillebrand, analyst.

RHE—Fire tower syenite, near fire tower at summit. F. A. Gonyer, analyst.

RHF—Watson Ledge quartz syenite, Watson Ledge. F. A. Gonyer, analyst.

RHG—Garland Peak syenite, top Garland Peak. F. A. Gonyer, analyst.

RHH—Interior fine granite. F. A. Gonyer, analyst.

Belknap Mountains (Modell, 1936)

BM3—Quartz diorite from Minge Cove, southeastern part of Winnipesaukee quadrangle. F. A. Gonyer, analyst.

BM4—Perthite granodiorite, west slope of Piper Mountain, near south edge of Winnipesaukee quadrangle H. S. Washington, analyst.

BM5—Gabbro, west foot of Locke's Hill. H. S. Washington, analyst.

BM6—Diorite, Locke's Hill. H. S. Washington, analyst.

BM7—Monzodiorite from Huckleberry Mountain. F. A. Gonyer, analyst

BM8—Syenite, west slopes of Mt. Belknap. H. S. Washington, analyst.

BM9—Albany porphyritic quartz syenite from near Ox Mountain. F. A. Gonyer, analyst.

BM10—Conway granite from the North Conway quadrangle; red phase, Redstone, New Hampshire. W. H. Herdsman, analyst.

Ossipee Mountains (Kingsley, 1931)

OM1—Basalt, Bald Knob, Ossipee Mountains. F. A. Gonyer, analyst.

OM2—Andesite, Cold Brook, Ossipee Mountains. F. A. Gonyer, analyst.

NEW HAMPSHIRE MAGMA SERIES

Littleton-Moosilauke Area (Billings, 1937)

LM13—Bethlehem gneiss, village of Easton, Moosilauke quadrangle. F. A. Gonyer, analyst.

LM15—Bethlehem gneiss, 0.2 mile S. 45°W. of Gilmanton School, Littleton quadrangle. F. A. Gonyer, analyst.

LM16—Kinsman quartz monzonite, nonporphyritic type, Kinsman Notch, on highway half a mile east of B.M. 1814, Moosilauke quadrangle. F. A. Gonyer, analyst.

LM17—Kinsman quartz monzonite, porphyritic type, Kinsman Notch, outlet of Beaver Pond, Moosilauke quadrangle. F. A. Gonyer, analyst.

Mt. Cube Area (Hadley, 1942)

MtC 442—Quartz monzonite of Bethlehem gneiss, Indian Pond pluton, Quinttown-Orfordville road, 0.7 mile northwest of Quinttown corner, Mt. Cube quadrangle. A. Willman, analyst.

analysis of Conway granite from Conway, New Hampshire, has been used because the Conway granite from the Belknap Mountains has not been analysed.

The lines for the Red Hill rocks (short dashes) are somewhat more irregular, but analogous. It is especially noticeable that the nepheline-soldalite syenite causes irregularity in the Red Hill curves.

On the same diagram are plotted Daly's (1933, p. 17) average plateau basalt, Kennedy's (1933, p. 241) olivine basalt, and Kennedy's tholeiitic magma. None agrees well with the gabbro from the Belknap Mountains. Kennedy postulated olivine basalt magma as the parent of the alkalic rocks, but there is no indication in these curves that it is more probable than the tholeiitic magma.

The simplicity of these curves, particularly for the Belknap rocks, suggests that the changes may have been caused by simple addition or subtraction. It was decided, accordingly, to calculate subtractions and additions which could produce the changes. It is realized that individual rocks may not be on the main line of change and that the calculations are valid only in a general way.

Change by subtraction.—Table 5 gives the smallest amounts of material which could be subtracted to cause the changes indicated for the Belknap rocks and for the Red Hill rocks. These results may be obtained graphically, as suggested by Bowen (1928, p. 87, 88), or arithmetically. Several norms of this subtracted material are given also. Some were omitted because they are of such composition that no provision is made in the calculation of norms.

The chemical character of these calculated subtractions does not correspond to the minerals which might be withdrawn by crystal fractionation. Especially unlikely are the two changes in the Belknap rocks, from gabbro (BM5) to diorite (BM6) and from syenite (BM8) to quartz syenite (BM9). The other three subtractions agree more closely with minerals which might be withdrawn from these rocks, but the plagioclase in the norm of the change from diorite (BM6) to monzodiorite (BM7) is more sodic than would form in a diorite, and the norm of the change from quartz syenite (BM9) to granite (RHH) has some quartz and 33 per cent orthoclase. Bowen (1928, p. 85–91) has pointed out that normative minerals do not always agree with modal minerals, but this does not seem to be a sufficient explanation of these subtractions, for the analyses of minerals of the White Mountain magma series (*see* Table 6) given by Chapman and Williams (1935, Table 7) do not agree with the subtractions, either. Account should be taken also of the fact, which Chapman and Williams point out (1935, p. 522, 523), that the ferromagnesian minerals which might be withdrawn from a magma would be more magnesian than the corresponding mineral in the rock, and the plagioclase would be more calcic.

The composition of the subtracted material could be changed somewhat by subtracting more than the minimum, but that is a somewhat doubtful remedy quantitatively because with minimum subtraction the granite is only 6 per cent of the original material. Greater subtraction would reduce the granite to even less. The change caused by increasing the amount of subtracted material is illustrated by the change from syenite (BM8) to quartz syenite (BM9) (Table 7). These compositions do not correspond to the minerals to be expected either.

Thus, the calculated subtractions do not agree with the minerals which might be

TABLE 5.—*Smallest amount of material (in weight per cent) which can be subtracted*

TABLE 5.—*Smallest amount of material (in weight per cent) which can be subtracted*

To change:

	Belknap area										Red Hill area								
	Gabbro BM5	to	Diorite BM6	to	Monzodiorite BM7	to	Syenite BM8	to	Quartz syenite BM9	to	Granite RHH	Syenite RHA	to	Neph.-sod. syenite RHB	to	Neph.-sod. syenite RHC	to	Syenite RHE	to	Granite RHH
SiO_2		35.0		54.1		52.3		.0		64.2			65.2		62.9		56.1		30.0	
Al_2O_3		19.7		12.7		18.6		48.3		18.9			17.8		9.9		23.9		33.6	
FeO		21.7		11.8		12.4		18.7		3.7			3.0		10.2		2.9		6.4	
MgO		7.5		6.2		3.7		7.7		.0			.0		2.6		.1		.0	
CaO		14.8		9.2		9.0		9.9		1.7			1.2		5.9		.0		6.0	
Na_2O		.0		6.0		4.0		5.5		5.9			4.8		.0		11.4		11.2	
K_2O		1.3		.0		.0		9.9		5.6			8.0		8.5		5.6		12.8	
		100.0		100.0		100.0		100.0		100.0			100.0		100.0		100.0		100.0	
Amount remaining	100		58		28		17		16		6	100		30		20		8		6
Norms:																				
Quartz										1.7			2.3		11.3					
Orthoclase										33.4			47.3		50.0		33.4			
Albite				37.7		34.1				49.8			40.3				25.7			
Anorthite				7.8		32.5				8.3			3.6		1.9					
Nepheline				7.1													35.5			
Sodium metasilicate																	1.2			
Diopside				31.6		10.4							2.0		23.3					
Hypersthene						12.1				6.7			4.5		13.3					
Olivine				15.8		10.9											4.3			
				100.0		100.0				99.9			100.0		99.8		100.1			

TABLE 6.—*Chemical composition of the minerals of the White Mountain magma series*
(Chapman and Williams, 1935, Table 7)

	4	5	6	7	8	9
SiO_2	47.58	46.56	37.40	46.98	34.96	none
TiO_2	.37	1.83	3.20	1.49	none	48.90
Al_2O_3	1.16	2.48	12.34	1.29	none	none
Fe_2O_3	2.60	6.24	4.16	11.93	none	9.80
FeO	24.21	27.27	25.84	23.38	36.77	38.97
MnO	.59	1.08	1.24	.24	.52	.36
MgO	3.34	1.35	2.20	.13	27.04	2.03
CaO	18.80	6.15	9.72	1.91	none	n.d.
Na_2O	.47	4.17	1.80	8.90	n.d.	n.d.
K_2O	.21	1.37	1.36	2.74	n.d.	n.d.
H_2O+	.34	1.27	tr	1.10	n.d.	n.d.
H_2O-		n.d.	.60	none	—	—
	99.67	99.77	99.86	100.09	99.29*	100.06

4. Hedenbergite from syenite, east slope of knoll south of Burnside Brook at elevation 2160 feet, Percy quadrangle.
5. Hornblende from syenite, in Moore Brook at elevation 2000 feet, Percy quadrangle.
6. Hastingsite from quartz syenite, Jackson Falls, Jackson, North Conway quadrangle.
7. Riebeckite from granite, ½ mile N.E. of small swamp on Mill Mt., Percy quadrangle.
8. Hyalosiderite from gabbro, Black Cascade, Tripyramid Mountain.
9. Ilmenite from gabbro, Black Cascade, Tripyramid Mountain.

All analyses except specimen 6 by F. A. Gonyer. Specimen 6 apparently by W. H. Herdsman. (*See* Chapman and Williams, 1935, p. 512, Table 7.)

* Chapman and Williams give total for Number 8 as 100.58.

TABLE 7.—*Subtraction of more than the minimum to change syenite (BM8) to quartz syenite (BM9)*

	Minimum (9%) subtracted	50% subtracted	83% subtracted
SiO_2	.0	55.9	60.4
Al_2O_3	48.3	22.5	20.4
FeO	18.7	5.8	4.8
MgO	7.7	1.4	.9
CaO	9.9	3.0	2.4
Na_2O	5.5	5.0	5.0
K_2O	9.9	6.4	6.1
	100.0	100.0	100.0

extracted by crystal fractionation at the appropriate stage, and crystal fractionation alone cannot explain the Belknap series. Similarly, the calculated subtractions for the Red Hill series are unlikely.

Chapman and Williams (1935, p. 522–524) attacked the problem in a somewhat different way and came to a different conclusion. They write, "... fractional crystallization alone is qualitatively capable of producing all the rock types of the White Mountain magma series" (p. 526). To the writer, however, two objections invalidate their conclusions. (1) Their calculations are only partly of specific application to the White Mountain magma series. They start with a gabbro from Tripyramid

Mountain, and the subtracted minerals are derived from analyses of minerals from the White Mountain magma series. The calculations are general, however, in that the monzonite, syenite, and granite obtained by this subtraction are "typical" and not any particular rocks. This objection appears more serious upon examination of Figure 2, where it is seen that rocks from different areas might make curves more irregular than the curves here made by rocks from one area. With several analyses from two different areas, for instance, different pairs, each pair including a rock from each area, would give quite different results for calculations of this sort. Calculations of this type should, therefore, deal exclusively with particular analyses from the same magma series and only with rocks from the same center of igneous activity. (2) The minerals subtracted by Chapman and Williams are not entirely probable. (*See* Table 8.) The plagioclase subtracted in the change from monzonite to syenite is slightly more calcic than the plagioclase in the change from gabbro to monzonite. Furthermore, the plagioclase subtracted from syenite to form granite is An_{20}. Therefore, the plagioclase of the granite should be more sodic than An_{20}, but it is not. This is somewhat complicated by the fact that much of the plagioclase is in the microperthite, however. It also seems improbable that 16 per cent of pyroxene could be subtracted from monzonite in the change to syenite or that 13 per cent of hastingsite could be subtracted from syenite in the change to granite.

TABLE 8.—*Calculations for "Derivation of rock types from a Gabbro magma by fractional crystallization"*
(In weight per cent; by Chapman and Williams, 1935, Table 9)

	Gabbro	Monzonite	Syenite	Granite
SiO_2	48.04	58.00	63.60	71.40
TiO_2	2.56	1.38	.61	.55
Al_2O_3	20.10	16.10	16.20	14.32
FeO	8.21	6.45	4.05	2.10
MgO	4.68	2.03	.80	.21
CaO	11.52	7.82	3.13	2.10
Na_2O	3.01	4.60	6.04	3.31
K_2O	.79	3.69	5.66	5.78
	98.91	100.07	100.09	99.77

Minerals subtracted to change:

	Gabbro to Monzonite	Monzonite to Syenite	Syenite to Granite
Orthoclase	—	—	20.02
Albite	17.00	5.24	19.91
Anorthite	36.00	11.95	5.00
Pyroxene	10.00	16.00	—
Olivine	10.00	—	—
Ilmenite	4.50	2.00	—
Hastingsite	—	—	13.00
	77.50	35.19	57.93

Monzonite with 16 per cent pyroxene and syenite with 13 per cent hastingsite are not given in their Table 1.

Chapman and Williams also rejected crystal fractionation as the sole method of change for these rocks on a quantitative basis, even though they thought it qualitatively possible. They state,

> "An objection to the derivation of the various rock types of the White Mountain magma series by fractional crystallization is based on the belief that tremendous quantities of gabbro would be necessary to produce the syenites and the granites, and, thus, there should be large quantities of basic rocks beneath the White Mountain area. According to isostasy, such a region should be a lowland and not a mountainous area" (p. 524).

Consequently, they invoked some assimilation and pure melting. They also concluded that the plutonic rocks, the ring dikes, and the associated stocks may extend as much as 8 or 10 miles down (p. 525). However, recent seismological work by Leet (1940; 1942; personal communication) indicates the presence of a layer, probably basaltic, at a depth of about 9 miles under New England. Their quantitative objection appears, therefore, not to hold, for contrary to their topographic evidence, there is mafic rock at about the bottoms of the intrusive bodies. The writer, although in agreement with their final conclusion, disagrees with their reasons.

Change by addition.—An examination of the materials to be added to produce these changes reveals that they, too, are improbable. (*See* Table 9.) So far as they are known, the older rocks in the Winnipesaukee area do not have the compositions indicated by the calculated additions. To be sure, some of the calculated additions approximate the compositions of schist or granite, but the alkalies, particularly, are too abundant in the calculated additions. The high alkali content is one of the important characteristics of the White Mountain magma series, and it cannot be ignored. Any such peculiarity of an original magma would be lost by dilution if much different material were added. Therein lies one of the strong objections to large-scale assimilation by this magma.

The variation between the added materials at different stages is in itself an objection to addition, for, with assimilation or pure melting, the same type of material would be added throughout the process. At least, the added material would probably not vary so widely or unsystematically as do the calculated additions.

A further strong objection to addition as the sole method of origin of this series is the great amount of material which must have been added. To carry the series from gabbro (BM5) to granite (RHH), as calculated step by step, requires the total addition of about 42 times the original quantity of material. Calculated as a single change from gabbro (BM5) to granite (RHH), the amount required is 10 times the original magma. It seems improbable that a magma can assimilate so much material.

It is concluded that addition, either by assimilation or pure melting, cannot alone explain the variation in this series.

Conclusion.—Although neither crystal fractionation nor addition by assimilation or pure melting can have been the sole method of change in the White Mountain

TABLE 9.—*Smallest amount of material (in weight per cent) which can be added*

TABLE 9.—*Smallest amount of material (in weight per cent) which can be added*

To change:	Belknap area									Red Hill area							
	Gabbro BM5	to	Diorite BM6	to	Monzodiorite BM7	to	Syenite BM8	to	Quartz syenite BM9	to	Granite RHH	Syenite RHA	to	Nepheline-sodalite syenite RHB	to	Nepheline-sodalite syenite RHC	to	Syenite RHE	to	Granite RHH
SiO_2		73.0		59.7		64.0		68.9		83.8			25.0		58.8		68.8		85.3
Al_2O_3		12.8		22.5		20.3		16.4		8.0			18.0		22.7		14.0		8.7
FeO		.0		5.6		2.3		2.8		.0			22.0		2.4		3.5		1.0
MgO		1.4		.0		.0		.0		1.4			8.0		.0		.7		.7
CaO		1.0		2.9		.4		1.4		1.1			13.0		.6		4.5		.0
Na_2O		9.7		4.0		5.3		5.0		.0			14.0		9.5		.0		1.2
K_2O		2.1		5.3		7.7		5.5		5.7			.0		6.0		8.5		3.1
		100.0		100.0		100.0		100.0		100.0			100.0		100.0		100.0		100.0	
Amount	100		151		326		774		3,096		4,180	100		110		550		930		1,488
Norms:																				
Quartz		21.7		5.2		1.1		13.4		57.5							24.8		64.7	
Corundum				4.9		2.6													3.4	
Orthoclase		12.2		31.1		45.6		32.8		33.9					35.6		50.0		18.3	
Albite		54.0		34.1		44.5		42.4							33.0				10.0	
Anorthite				14.5		2.0		5.8		4.7					1.7		13.1			
Nepheline															25.8					
Sodium metasilicate		6.5																		
Diopside		3.9						1.0		.7					1.2		7.9			
Hypersthene		1.7		10.3		4.2		4.6		3.2							4.6		3.7	
Olivine															2.9					
		100.0		100.1		100.0		100.0		100.0					100.2		100.4		100.1	

magma series, probably both played a part. Crystal fractionation was probably the dominant process because: (1) The order of intrusion agrees with that expected in crystal fractionation, and (2) the high alkali content is characteristic of considerable range of rocks in this series. Whatever the cause of this alkalic character, it would be lost by dilution if much of the known available rocks were assimilated. The alkalic character of the White Mountain magma series is most pronounced at about the syenite to feldspathoid stage and tends to become less marked toward the granite stage (Figs. 2, 3). Subtraction by crystal fractionation appears most unlikely in the change from syenite to granite. Addition at the same stage is more reasonable, although some of the material to be added is more siliceous than most of the known rocks here. As suggested by Modell (1936, p. 1916), "The problem of differentiation is chiefly concerned with the production not of syenite, but of granite".

The intrusions, therefore, probably formed by crystal fractionation and assimilation combined, with crystal fractionation dominant up to the stage of syenite and with assimilation taking on increased importance in the change to granite.

Both the subtractions and the additions in the part of the Red Hill series which involves the nepheline-sodalite syenite are improbable, indicating that the feldspathoidal syenite needs some unusual, perhaps complex, explanation. For instance, there was enough soda in the magma to cause the magmatic crystallization of some nepheline, but some of the sodalite was formed by later soda-bearing solutions (Quinn, 1937, p. 386). This rock is considerably off the main trend (Fig. 2), which is further indication of its unusual character.

The presence of the earlier unsystematic extrusions in the Ossipee Mountains and in several other areas of the White Mountain magma series is not yet explained.

SUMMARY AND CONCLUSIONS

The oldest rocks of this area are schist correlated on the basis of lithology, metamorphism, and structure with the early Devonian Littleton schist.

Intrusive into the schist is the New Hampshire magma series, including the Meredith porphyritic granite, the Winnipesaukee quartz diorite, hornblende gabbro, and Concord granite. The Meredith porphyritic granite with its large crystals of potash feldspar was intruded as a liquid with the feldspar crystals and numerous inclusions of schist floating in it. The evidences for this conclusion are the large size of the body, the sharp contacts of the schist inclusions, the textural evidence of early crystallization of the potash feldspar, and the alignment of the phenocrysts parallel to the contacts.

The New Hampshire magma series is characterized by the absence of extrusions, by large intrusions, weak foliation, a partially concordant relation to the schist, "normal" content of alkalies, a small compositional variation within a restricted range, the unsystematic order of intrusions, and the scarcity of associated basaltic or gabbroic rocks. No definite conclusion about the origin of this series is drawn, but origin by melting of the sial is considered.

The youngest series of rocks, the White Mountain magma series, is represented by numerous extrusions and intrusions. This series is characterized by early unsystematic extrusions, small discordant intrusions in the form of ring dikes and stocks,

a high alkali content, the presence of gabbroic and basaltic rocks, and the systematic order of intrusions from gabbro to granite.

Calculations are made to determine the material which could be subtracted or added to form the intrusive rocks. These calculations indicate that neither subtraction by crystal fractionation nor addition by assimilation or pure melting can alone explain the variations. Both were probably operative, with crystal fractionation more important in the early stages up to the formation of syenite and with addition dominant in the later stages. The presence of the early unsystematic extrusions has not been satisfactorily explained.

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PETROLOGY AND STRUCTURE OF THE OLIVERIAN MAGMA SERIES IN THE MOUNT WASHINGTON QUADRANGLE, NEW HAMPSHIRE

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CONTENTS

	Page
Abstract	497
Introduction	499
Acknowledgments	499
General lithologic features	500
Albee formation	500
Ammonoosuc volcanics	500
Partridge formation	501
Oliverian magma series	501
General statement	501
Hornblende-quartz monzonite	503
Fine-grained gray quartz monzonite	503
Coarse syenite	504
Coarse granite	505
Biotite-quartz monzonite	505
Porphyritic biotite-quartz monzonite	506
Pegmatite and aplite	507
Comparative mineralogy	507
Consanguinity of the group	508
Age relations	508
Origin	511
Structure	512
General relations	512
Structure of surrounding metamorphic rocks	512
Structure of igneous core	513
Nature of intrusion	513
Correlation with Oliverian magma series elsewhere	514
Summary	515
References cited	516

ILLUSTRATIONS

Figure	Page
1. Index map of New Hampshire	498
2. Ranges in composition of plagioclase and biotite	509
Plate	**Facing page**
1. Geological map and cross section of part of the Mount Washington quadrangle, New Hampshire	516

ABSTRACT

Plutonic rocks of four magma series occur in the Mt. Washington quadrangle, northern New Hampshire, but those of the Oliverian magma series are most abundant. In this quadrangle this late Devonian (?) series intrudes the Ordovician (?) Ammonoosuc volcanics and consists of six map-

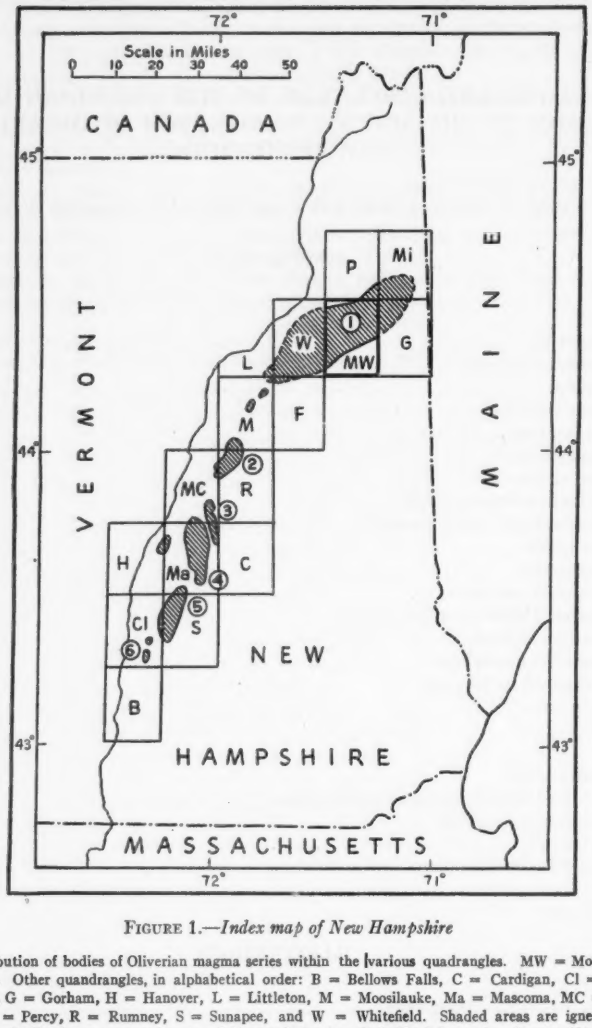

FIGURE 1.—*Index map of New Hampshire*

Shows distribution of bodies of Oliverian magma series within the various quadrangles. MW = Mount Washington quadrangle. Other quandrangles, in alphabetical order: B = Bellows Falls, C = Cardigan, Cl = Claremont, F = Franconia, G = Gorham, H = Hanover, L = Littleton, M = Moosilauke, Ma = Mascoma, MC = Mt. Cube, Mi = Milan, P = Percy, R = Rumney, S = Sunapee, and W = Whitefield. Shaded areas are igneous cores of principal domes of Oliverian magma series: 1 = Jefferson dome; 2 = Owls Head dome; 3 = Smarts Mtn. dome; 4 = Mascoma dome; 5 = Croydon dome; 6 = Unity dome. Domes south of Claremont quadrangle are not shown.

pable units: hornblende-quartz monzonite, fine-grained gray quartz monzonite, porphyritic biotite-quartz monzonite, biotite-quartz monzonite, coarse syenite, and coarse granite. Although most of these types occur in the Oliverian magma series elsewhere in New Hampshire, the coarse syenite and hornblende-quartz monzonite appear to be unique.

Potash feldspar in these rocks is generally microcline and uncommonly orthoclase; perthitic intergrowths are rare. Oligoclase is the characteristic plagioclase, except in the hornblende-quartz mon-

zonite, in which it is andesine. Biotite, the most abundant dark mineral, has a considerable range in chemical composition assuming that the γ index is a reliable indicator. The iron-magnesia ratio of the biotite is considerably higher in the hornblende-free rocks than in those containing hornblende. The amphibole in this magma series is common hornblende, and the muscovite contains some iron and magnesia.

The igneous rocks form the core of the Jefferson dome, which, essentially concordant, is more than 40 miles long and 9 miles wide. In most respects this dome resembles those in which the Oliverian magma series is found elsewhere in New Hampshire. It differs principally in its larger dimensions and composite character. The various petrographic units were intruded separately as large lenses, each developing planar flow structures that are shown by oriented minerals and xenoliths. The texture is commonly granoblastic.

INTRODUCTION

This paper is the third of a series presenting the results of a detailed study of the bedrock geology of the Mt. Washington quadrangle of northern New Hampshire (Fig. 1). The first paper, by Billings (1941), was concerned primarily with the structure and metamorphism of the southern part of the quadrangle. The second paper, by R. W. Chapman (1942), deals with the White Mountain magma series in the northern half of the quadrangle. The present paper, which describes the Oliverian magma series, is concerned with problems only briefly mentioned in the other two papers, although geographically the three reports overlap somewhat. An earlier publication (R. W. Chapman, 1937) describes the small stock of the White Mountain magma series at Cherry Mountain.

The Oliverian magma series of the Mt. Washington quadrangle is worthy of special description because of its variability and its numerous lithologic types. Whereas the rocks of the Oliverian magma series elsewhere in New Hampshire are relatively uniform, in the Mt. Washington quadrangle they are very variable. Moreover, these rocks constitute the largest pluton of this magma series in New Hampshire. In fact, the limits of this pluton are far beyond the borders of the Mt. Washington quadrangle, in areas where the geology has not yet been mapped. A complete structural study of the pluton cannot be made until this mapping has been completed. The principal purpose of the present paper, therefore, is to describe the lithology of the Oliverian magma series and to consider briefly the structure and the mechanics of intrusion.

The field work consumed the entire summers of 1938 and 1939, and a month during each of the summers of 1936 and 1940. Photostats of the United States Geological Survey topographic sheet (1936), enlarged to a scale of 3 inches to the mile, were used for base maps. Most outcrops were located by means of aneroid barometer, on the assumption that the topographic map, surveyed in 1934–1935, was essentially correct. Billings is responsible for the mapping (Pl. 1) south of a line following the highway through Jefferson, Jefferson Highland, and Bowman, thence northward to Mt. Randolph, thence eastward to the east border of the map. The Chapmans are responsible for the mapping north of this line.

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GENERAL LITHOLOGIC FEATURES

Metamorphic rocks belonging to the Albee formation, Ammonoosuc volcanics, and Partridge formation, all Ordovician (?), underlie a small part of the area considered in this report.

The igneous rocks of the area belong to four magma series: (1) Highlandcroft, (2) Oliverian, (3) New Hampshire, and (4) White Mountain.

The Highlandcroft magma series is the oldest and is represented by the Lost Nation group (R. W. Chapman, 1935, p. 405) in the northwest corner of the area (Pl. 1) The group is essentially quartz diorite and gabbro-diorite, rather intensely metamorphosed and in most places distinctly foliated. This series has no direct bearing on the major problems of this paper.

The Oliverian magma series is the next oldest, and consideration of it constitutes the major part of this paper. The rocks of this series form a gigantic dome, which, in the Mt. Washington quadrangle, is bounded on the northwest by a discontinuous belt of the Ammonoosuc volcanics and on the southeast by a somewhat discontinuous belt of the Ammonoosuc volcanics and Partridge formation. The series includes principally granite, quartz monzonite, granodiorite, and syenite. The rocks are younger than the Ammonoosuc volcanics but older than the late Devonian folding.

The New Hampshire magma series is represented chiefly by stock-like masses of Bickford granite which Billings (1941, p. 898) has shown were intruded after the late Devonian folding. Like the Lost Nation group, the Bickford granite has no direct bearing on the problems of this paper.

The youngest intrusives belong to the White Mountain magma series and occur principally in the northern part of the area as subcircular stocks and ring-dikes (R. W. Chapman, 1942). This alkaline series is chiefly granite, quartz syenite, syenite, and quartz monzodiorite. These rocks are intrusive into the Oliverian magma series in this area. Because the White Mountain magma series is not discussed in this paper, the various petrographic types are not distinguished on the map (Pl. 1).

ALBEE FORMATION

Exposure of the Albee formation (Upper Ordovician?) are confined to the northwestern part of the Mt. Washington quadrangle. One set of outcrops occupies an area 1 mile long and 1000 feet wide 4½ miles north of the village of Jefferson. Numerous inclusions of this formation, nine of which are shown on Plate 1, occur in the White Mountain magma series near the northern border of the quadrangle.

The Albee formation in this area consists of light-gray to dark-gray, fine-grained quartzite, locally stratified. Correlation with the Albee formation in the type locality is based on lithologic similarity. Quartzite, exposed a few miles to the north, in the Percy quadrangle, has been correlated (R. W. Chapman, 1935, p. 405) with the Albee formation of the type locality in the Littleton-Moosilauke area (Billings, 1937, p. 472). Reconnaissance indicates that the quartzites of the Percy area can be traced continuously to the type locality.

AMMONOOSUC VOLCANICS

A somewhat discontinuous belt of Ammonoosuc volcanics (Upper Ordovician?) trends northeast along the southern boundary of the area. Billings (1941, p. 872)

has already described this formation, and only the essential features will be given here. The volcanics are chiefly biotite gneiss, but small amounts of amphibolite, quartz-muscovite-pyrite schist, and lime-silicate granulite are found. The biotite gneiss is a gray, fine-grained rock composed essentially of quartz, plagioclase, potash feldspar, and biotite. It is believed to have formed by metamorphism from pyroclastic rocks ranging from rhyolite to dacite. Amphibolite, a dark, medium-grained rock composed of hornblende and andesine, was probably derived from basaltic or andesitic material. Gray to green lime-silicate granulites, generally composed of quartz, andesine, hornblende, diopside-hedenbergite, epidote, and chlorite, are believed to have formed from basalt and quartz latite. The quartz-muscovite-pyrite schist probably represents argillaceous sandstone interbedded with the volcanics.

The Ammonoosuc volcanics also occupy several square miles in the northwest corner of the Mt. Washington quadrangle. Immediately east of these outcrops the bedrock is hidden by glacial drift, and still farther east and northeast the intrusives of the White Mountain magma series lie on the strike of the volcanic rocks. Reconnaissance shows that the Ammonoosuc volcanics extend for several miles to the west into the Whitefield quadrangle.

These volcanics, like those in the southern part of the quadrangle, are largely fine-grained biotite gneiss, with minor amounts of amphibolite and mica schist. Large knots and lenses of epidote-rich material, up to a foot or so across, are abundant in many outcrops. Such features are characteristic of the Ammonoosuc volcanics elsewhere.

A pyroclastic origin for some of these rocks is indicated in many places. In a well-preserved breccia at the Gore School, $3\frac{1}{4}$ miles northwest of the summit of Mt. Starr King, a matrix of biotite gneiss contains closely packed blocks of fine-grained gneiss and amphibolite. Elsewhere, subhedral grains of feldspar and quartz, like those in crystal tuffs, are common.

The Ammonoosuc volcanics exposed in the southern part of the quadrangle are about 5000 feet thick (Billings, 1941, p. 877). The base of the formation, however, is nowhere exposed, due to its intrusive relationship with the rocks of the Oliverian magma series.

PARTRIDGE FORMATION

The Partridge formation (Upper Ordovician?) overlies the Ammonoosuc volcanics in the southern part of the area. Billings (1941, p. 877) has described these rocks in detail. Medium-grained to fine-grained gneiss, derived from shale, constitutes most of the formation. It is composed of alternating dark and light layers ranging in thickness from 0.5 to 3 centimeters. Biotite, muscovite, and quartz make up the dark layers, whereas quartz, andesine, and small quantities of biotite and muscovite compose the light layers. The Partridge formation is approximately 1400 feet thick (Billings, 1941, p. 881).

OLIVERIAN MAGMA SERIES

GENERAL STATEMENT

The rocks of the Oliverian magma series in the Mt. Washington quadrangle are plutonic, showing a considerable range in composition, texture, and structure. They are composed largely of potash feldspar, plagioclase, and quartz, with subordinate

TABLE 1.—*Modes of the Oliverian magma series*

	1		2		3		4		5		6		7
Quartz	14	4–20	28	25–30	6	tr–19	33	27–39	33	16–40	31	28–36	30
Potash feldspar	37	23–60	24	17–35	59	31–76	43	31–54	33	19–42	34	20–40	52
Plagioclase	31	18–40	36	24–49	29	15–50	20	10–30	29	18–42	27	20–43	10
Muscovite			1.5	0–2					1	0–2	2	0–5	
Biotite	5	2–10	10	6–20	2	tr–6	2	1–3	4	2–7	5	0–7	
Chlorite	tr	0–tr	tr	0–tr	tr	0–1	tr	0–1	tr	0–1	1	0–5	1
Hornblende	12	6–18			3.5	0–9	1	0–2					4
Magnetite	tr	0–10	1	tr–2	0.5	tr–2	1	tr–1	0.5	0–2	0.5	0–2	1
Sphene	1	tr–2			tr	tr–1	0.5	tr–1	tr	0–tr	tr	0–tr	tr
Apatite	tr	tr	tr	tr	tr	tr	tr	0–tr	tr	0–tr	tr	0–tr	tr
Zircon	tr	0–tr	tr	0–tr	tr	tr	tr	0–tr	tr	0–tr	tr	0–tr	tr
Allanite	tr	0–tr	tr	0–tr	tr	0–tr	tr	0–tr	tr	0–tr			
Epidote	tr	0–tr	tr	0–tr	tr	0–1	tr	0–tr	tr	0–tr	tr	0–tr	2
Pyroxene					tr	0–tr	tr	0–tr					
Garnet									tr	0–tr			
Rutile	tr	0–tr	tr	0–tr	tr	0–tr			tr	0–tr			
Tourmaline					tr	0–1							
Andalusite											tr	0–1	
Sillimanite											tr	0–1	
Composition of plagioclase (% An)	33	21–42	22	16–32	21	14–37	18	15–20	17	13–23	17	14–20	
Grain size in mm (average range)	0.2–9.0		0.2–2.0		0.2–12.0		0.4–10.0		0.2–3.0		0.2–9.0		0.1–0.4

1. Hornblende-quartz monzonite, 6 sections.
2. Fine-grained gray quartz monzonite, 4 sections.
3. Coarse syenite, 11 sections.
4. Coarse granite, 6 sections.
5. Biotite-quartz monzonite (biotite gneiss), 10 sections.
6. Porphyritic biotite-quartz monzonite (porphyritic biotite gneiss), 8 sections.
7. Hornblende granite dike, 1 section.

Under each lithologic type the left-hand column is the average, the right-hand column is the range.

biotite, hornblende, and accessories. Most varieties are medium-grained or coarse-grained, but fine-grained phases are locally abundant. A foliation, formed by parallel arrangement of mafic minerals, feldspar, and quartz, is generally megascopically noticeable. Some phases, however, are typically massive, whereas others are strongly schistose. A number of petrographic types are recognized. These, in general, occur in more or less distinct units and have been mapped separately. Locally some types are transitional, and elsewhere two or more types are intimately associated. The distribution of the various rock types is given in Plate 1.

The following types are separately mapped and described: (1) hornblende-quartz monzonite, (2) fine-grained gray quartz monzonite, (3) coarse syenite, (4) coarse granite, (5) biotite-quartz monzonite, (6) porphyritic biotite-quartz monzonite.

Under each lithologic type the left-hand column is the average, the right-hand column is the range.

HORNBLENDE-QUARTZ MONZONITE

Hornblende-quartz monzonite is exposed near the main highway, between the villages of Jefferson Highland and Jefferson. All gradations are found between medium-grained to coarse-grained porphyritic varieties and medium-grained to fine-grained nonporphyritic varieties.

The porphyritic variety is well exposed half a mile southwest of Jefferson Highland (in Stag Hollow Brook, altitude of 1220–1352 feet). Pink to white tabular phenocrysts of potash feldspar, 5 to 30 millimeters long, composed of Carlsbad twins, are set in a dark-gray groundmass of plagioclase, potash feldspar, quartz, hornblende, and biotite. Locally the rock shows pronounced planar structure, produced by the alternating bands of dark and light constituents and the parallelism of tabular crystals of potash feldspar. The porphyritic variety is also exposed locally along the main highway between Jefferson Highland and Jefferson.

With diminution in the size and number of the phenocrysts of potash feldspar, the porphyritic variety passes into the nonporphyritic variety. The nonporphyritic variety resembles some of the hornblendic rocks of the Ammonoosuc volcanics.

Microscopic study of the hornblende-quartz monzonite shows a granoblastic texture. In the porphyritic variety potash feldspar (both orthoclase and microcline) is present in about the same proportions as plagioclase, and the rock is, therefore, a quartz monzonite. In the nonporphyritic variety, however, the ratio of potash feldspar to plagioclase is considerably less, and the rock approaches a granodiorite. The plagioclase averages sodic andesine (range from An_{21} to An_{42}); this is more calcic than the plagioclase of any of the other petrographic units. Olive-green to brown biotite has an average γ index of 1.637, indicating a lower iron-magnesia ratio than in any of the other petrographic units; the index suggests about 50 per cent annite-siderophyllite, 50 per cent phlogopite-eastonite. The γ index of the hornblende ranges from 1.674 to 1.688, and $2V = 55°–65°$. In one specimen $\alpha = 1.660$, $\beta = 1.672$, and $\gamma = 1.681$. These data are similar to those obtained from an analysed common hornblende in the Littleton-Moosilauke area (Billings, 1937, p. 513, 556). An average mode of the hornblende-quartz monzonite is given in Table 1.

FINE-GRAINED GRAY QUARTZ MONZONITE

This type is confined to relatively small areas, two of which are shown on Plate 1: a mile northeast of Bowman and three quarters of a mile southeast of the summit of Mt. Randolph. The rock is one of the finest-grained types in the Oliverian magma series of this area. Gray and relatively massive, it locally possesses a fair foliation.

Microscopic study shows that the texture is granoblastic and that the essential minerals are oligoclase (An_{22}), potash feldspar, quartz, and biotite. Accessories are muscovite and magnetite. The average γ index of the biotite is 1.652, suggesting about 60 per cent annite-siderophyllite. The muscovite has an average γ index of 1.615, and $2V = 35°$; these data suggest that some iron and magnesia are present.

COARSE SYENITE

One of the most abundant rock types of the Oliverian magma series in the Mt. Washington quadrangle is coarse syenite, the mapped area of which is about 35 square miles. Were it not for intrusives of the White Mountain magma series, which, in large part, have taken the place of the older syenite, the coarse syenite would cover an area of about 55 square miles. Outcrops of typical syenite are most accessible along the north side of the main highway, half a mile northwest of Jefferson village. Elsewhere in New Hampshire quartz-poor rocks are rare in the Oliverian magma series. Exclusive of the Mt. Washington quadrangle, about 150 square miles of rocks of this magma series have been mapped in New Hampshire, and they all range from quartz diorite to granite. The quartz content is rarely less than 20 per cent and the range is generally from 25 to 45 per cent. The coarse syenite, therefore, constitutes an unusual petrographic type for the Oliverian magma series.

The coarse syenite is pink, locally medium-grained, and composed predominantly of large crystals of potash feldspar. Planar structure is common. There are all gradations between coarse, porphyritic phases and medium-grained equigranular phases. In the coarse phase, pink crystals of potash feldspar, 5 to 30 millimeters long, are commonly arranged in trachytoid fashion, and in places the rock has a fair foliation. The feldspar is subhedral to euhedral and usually shows Carlsbad twinning. These large crystals constitute most of the rock and are set in a gray to black groundmass, which in most instances is nearly aphanitic. Biotite, in tiny flakes, is about the only megascopically visible mineral of the groundmass. In an outcrop half a mile northwest of Jefferson village, feldspar crystals up to 7.5 centimeters long were observed. The equigranular phase is more commonly medium-grained and seldom foliated, and pink to white potash feldspar occurs in nearly equidimensional grains averaging 5 millimeters across. Smaller quantities of quartz, biotite, hornblende, and plagioclase may be detected with the hand lens. About 3 miles north of Jefferson village, the rock contains as much as 20 per cent quartz and is strictly a granite. This quartz-rich phase is believed, however, to be a local variety of the syenite.

Microscopic study of the porphyritic phase shows that the large crystals of potash feldspar are euhedral to anhedral microperthite; the host of the perthite seldom shows microcline twinning. Considerable potash feldspar occurs in small anhedral grains between the large crystals. In the nonporphyritic phase the potash feldspar is also perthite, rarely showing microcline twinning, and the texture is allotriomorphic granular. The plagioclase is usually oligoclase but ranges from An_{14} to An_{27}. Generally in anhedral grains, it shows albite twinning and occasionally also Carlsbad or pericline twinning. Locally myrmekite replaces potash feldspar in considerable quantities. Quartz is most commonly interstitial to perthite and in some instances poikilitically encloses tiny grains of plagioclase. Green and brown flakes of biotite are usually segregated with other mafic minerals and accessories. The γ index averages 1.640, which indicates biotite of intermediate composition. The optical data for the hornblende are as follows: γ ranges from 1.677 to 1.691; $Z \wedge c$ averages 21°; and $2V$ ranges from 50° to 70°. These data indicate common hornblende. Pyroxene is so rare that it

was impossible to obtain optical data. Perhaps the most characteristic accessory is zircon, which is abundant, commonly in crystals up to 0.2 millimeter long. Although present in the other rocks of the Oliverian magma series, it appears to be most typically associated with the coarse syenite. An average mode of the coarse syenite is given in Table 1.

COARSE GRANITE

Coarse granite occupies a large area several miles wide, extending northeasterly from Cherry Mountain for about 10 miles. Exposures of typical material may be seen along the main highway, $2\frac{1}{2}$ miles southeast of Jefferson Highland.

The granite, in most places pink and coarse-grained, is composed of euhedral to subhedral crystals of potash feldspar 5 to 10 millimeters long, aggregates of milky to smoky quartz up to 10 millimeters long, and white granular aggregates of plagioclase 2 to 5 millimeters long. A massive structure prevails, but locally a crude planar structure is shown by the parallel arrangement of biotite flakes or tabular crystals of potash feldspar. Locally the rock is subporphyritic with phenocrysts of pink potash feldspar in a groundmass of finer-grained quartz, plagioclase, and biotite. This subporphyritic phase is identical with the Scrag granite of the Littleton, Moosilauke, and Franconia quadrangles (Billings, 1937, p. 510; Williams and Billings, 1938, p. 1024); the coarse granite of the Mt. Washington quadrangle and the Scrag granite to the southwest probably form a continuous belt across the southern part of the Whitefield quadrangle (Fig. 1).

On the average (Table 1) potash feldspar is distinctly dominant over plagioclase, and the rock is a true granite. The potash feldspar is chiefly microcline, for it shows the typical cross-hatched twinning, and locally microcline-perthite is abundant. Myrmekite replaces microcline in some specimens. Carlsbad twins are very conspicuous both under the microscope and in the hand specimen. The plagioclase, which is oligoclase averaging An_{18}, is in subhedral grains, from 0.5 to 5.0 millimeters across, and shows albite twinning. The quartz shows strain shadows, and many aggregates apparently represent large granulated grains. Biotite is in small grains, more or less grouped in aggregates, and averages 2 per cent of the rock. The γ index averages 1.642, suggesting biotite of intermediate composition. Common hornblende constitutes about 1 per cent of the rock and has the following optical properties: biaxially negative; 2V ranges from 50° to 60°; X = yellow, Y = green, Z = blue-green; $X<Y=X$; $X=b$, $Z \wedge c = 19°$; dispersion medium, $r>v$. The indices vary somewhat in different specimens; γ ranges from 1.669 to 1.675. For one specimen the following data were obtained: $\gamma = 1.675$, and $\beta = 1.668$; the birefringence is about 0.020, hence α would be $1.655\pm$. These data are similar to those for an analyzed common hornblende from an amphibolite in the Littleton-Moosilauke area (Billings, 1937, p. 513, 556). Pyroxene was observed in two thin sections, but in such small amounts that detailed optical data could not be obtained.

BIOTITE-QUARTZ MONZONITE

Biotite-quartz monzonite, called biotite gneiss in previous publications (Billings, 1941; R. W. Chapman, 1942), forms a long belt trending northeasterly from Cherry Mountain to the northeast corner of the area. Exposures are most numerous on the south slopes of Mt. Randolph and Mt. Crescent.

Megascopically the biotite-quartz monzonite is a pink, medium-grained rock characterized by granular texture and scarcity of dark minerals. Foliation is weak in some localities, but in many places it is pronounced. Generally the foliation is due to parallel orientation of small biotite flakes, but in places, as on the southeast slope of Mt. Crescent and Mt. Randolph, it is accentuated by large quartz and feldspar grains that have been flattened into thin lenses. In some outcrops the de-

formed quartz grains show a distinct lineation. The pink color of the rock is due to the abundance of potash feldspar. Locally potash feldspar occurs as small phenocrysts. Such rocks are intermediate in character between the biotite-quartz monzonite and the porphyritic biotite-quartz monzonite.

Microscopically the biotite-quartz monzonite shows a granoblastic texture. An average mode (Table 1) shows that potash feldspar and plagioclase are roughly in equal proportions; therefore the rock is a quartz monzonite. Some specimens, however, are granite, whereas others are granodiorite; but, inasmuch as the various phases of the biotite-quartz monzonite are transitional and very irregularly distributed, they were not mapped separately. In the northeastern corner of the map area the rock is poor in potash feldspar and is locally quartz diorite. This belt continues north, into the Percy quadrangle, where it has been described by R. W. Chapman (1935, p. 405) as a typical trondhjemite. Here the rock was given the name Berlin gneiss and tentatively assigned to the New Hampshire magma series. The present study shows, however, that the Berlin gneiss belongs to the Oliverian magma series.

The potash feldspar is predominantly microcline and in many cases shows Carlsbad twinning. Perthitic streaks may be observed in many grains, but the amount of included albite is usually small. Myrmekite replaces potash feldspar in small quantities in some specimens. The plagioclase is oligoclase (An_{12} to An_{23}), characterized by albite twinning, and some crystals show pericline or Carlsbad twins. Quartz characteristically shows undulatory extinction, indicating some deformation after crystallization. The amount of quartz is relatively constant throughout the rock and averages 33 per cent. Olive-green to olive-brown biotite, in places slightly chloritized, averages 4 per cent of the rock. Usually it occurs in small isolated flakes, many of which contain minute inclusions of zircon with pleochroic halos. The γ index for the biotite is 1.659, suggesting about 70 per cent annite-siderophyllite. Muscovite is rare in the Oliverian magma series but occurs in small quantities in the biotite-quartz monzonite in the Mt. Washington area. It commonly replaces biotite or occurs in small flakes parallel to the albite and pericline twin planes of the plagioclase. The average γ index of muscovite, which constitutes only 1 per cent of this rock, is 1.612, and $2V$ averages 33.5°. These data indicate that the muscovite contains some iron and magnesia. The most common minor constituents are magnetite, apatite, and zircon. Other minor constituents are shown in Table 1.

PORPHYRITIC BIOTITE-QUARTZ MONZONITE

The porphyritic biotite-quartz monzonite, called porphyritic biotite gneiss in previous publications (Billings, 1941; R. W. Chapman, 1942), is gradational into the biotite-quartz monzonite, but it was possible to map these two phases separately. The porphyritic phase, much less abundant than the nonporphyritic phase in the Mt. Washington quadrangle, is confined principally to the southwestern part of the area. In addition two small areas, on the south slope of Mt. Crescent, are shown in Plate 1.

These pink to gray rocks resemble the biotite-quartz monzonite but carry phenocrysts of pink or white potash feldspar, which are as much as 15 mm. long and commonly show Carlsbad twins. The medium-grained to fine-grained groundmass composed of feldspar, quartz, and mica, shows poor to fair foliation. Near Mt. Crescent the feldspar phenocrysts are more or less lens-shaped, and their major axes are in the plane of foliation. In some specimens the lenses are elongated parallel to a marked lineation in the rock. In general the foliation wraps around the large crystals, which appear to have been somewhat granulated.

Under the microscope the porphyritic biotite-quartz monzonite shows a granoblastic texture. The essential minerals are quartz, potash feldspar, and oligoclase, with minor amounts of biotite, muscovite, chlorite, and magnetite. An average mode is given in Table 1. The porphyritic biotite-

quartz monzonite, ranging from granite to granodiorite, averages quartz monzonite. The plagioclase is oligoclase (An_{14}–An_{20}). The γ index of the biotite averages 1.649, which suggests about 65 per cent of the annite-siderophyllite molecule. The γ index of muscovite averages 1.609; 2V averages 34°. These data indicate that the muscovite contains some iron and magnesia.

PEGMATITE AND APLITE

Pegmatites and aplites are commonly associated with some of the petrographic units. They are perhaps most abundant along the southeastern margin of the area occupied by the Oliverian magma series and in the overlying Ammonoosuc volcanics. They occur principally in the biotite-quartz monzonite and porphyritic biotite-quartz monzonite and to a less extent in the coarse granite. No attempt was made to study these bodies in detail. They are characteristically pink and composed of microcline, quartz, biotite, and muscovite. At several localities the bodies are composed in part of graphic granite.

The aplites are usually sill-like pink to gray, massive to weakly foliated, with a very small percentage of biotite and muscovite.

COMPARATIVE MINERALOGY

Having completed the petrographic descriptions of the Oliverian magma series, it is now appropriate to consider the range in composition of each mineral species throughout the magma series. The more abundant minerals in the various rock units are potash feldspar, plagioclase, quartz, biotite, hornblende, and muscovite. Subordinate constituents are magnetite, sphene, apatite, zircon, allanite, chlorite, epidote, pyroxene, garnet, rutile, tourmaline, andalusite, and sillimanite.

Potash feldspar, generally in flesh-colored subhedral grains or phenocrysts, is generally microcline and uncommonly orthoclase. Perthitic intergrowths with plagioclase are not common. Considerable microcline is in small grains in granoblastic development with quartz and plagioclase.

The range of plagioclase in the various rock units is shown in Figure 2. The determinations are based on index of refraction curves (Goranson, 1926, p. 140). The curves show that the anorthite content ranges from 10 to 42 per cent. In the hornblende-quartz monzonite the plagioclase is andesine. In the coarse granite, biotite-quartz monzonite, and porphyritic biotite-quartz monzonite it is oligoclase. In the coarse syenite and fine-grained gray quartz monzonite the plagioclase is chiefly oligoclase, but in some rocks it is andesine.

Biotite is biaxial negative with an optic angle ranging from 0 to 12°. The average is about 10° in the coarse granite; 8° in the biotite-quartz monzonite, porphyritic biotite-quartz monzonite, fine-grained gray quartz monzonite, and hornblende-quartz monzonite; and 5° in the coarse syenite. The maximum range for each rock unit is usually about 3°, except the porphyritic biotite-quartz monzonite, where the range is 12°. X = yellow, Y and Z = olive-green to olive-brown. Red-brown biotite was observed only in three specimens of the porphyritic biotite-quartz monzonite. In all specimens $X<Y = Z$. The ranges in composition of the biotite, as determined by the γ index (Winchell, 1933, p. 274), are shown in Figure 2. These curves show that the biotite of the hornblende-quartz monzonite, coarse syenite, and most of the coarse granite is much lower in annite-siderophyllite than that of the biotite-quartz monzonite, fine-grained gray quartz monzonite, and most of the por-

phyritic biotite-quartz monzonite. Though γ ranges from 1.635 to 1.664, it is distinctly lower in the hornblende-bearing rocks.

Hornblende is not common in the Oliverian magma series elsewhere in New Hampshire but is widely distributed through these rocks in the Mt. Washington quadrangle. It is biaxial negative, $2V = 50°–70°$, and the dispersion is moderately strong with $r > v$. X = yellow, Y = green, Z = greenish blue, and $X < Y = Z$. The γ indices are somewhat variable but are generally lower in the coarse granite (average 1.673, range 1.669–1.675) and higher in the hornblende-quartz monzonite, coarse syenite, and fine-grained gray quartz monzonite (average 1.683, range 1.674–1.691). The optical properties of several specimens were close to those of an analyzed common hornblende from the Littleton-Moosilauke area (Billings, 1937, p. 516).

Muscovite, as is to be expected, is absent from the hornblende-bearing rocks. The average content in those rocks that contain some muscovite is about 2 per cent. The γ index averages 1.612 in the biotite-quartz monzonite, 1.609 in the porphyritic biotite-quartz monzonite, and 1.615 in the fine-grained gray quartz monzonite. 2V averages 33.5° in the biotite-quartz monzonite, 34° in the porphyritic biotite-quartz monzonite, and 34.5° in the fine-grained gray quartz monzonite. These relatively uniform data indicate that the muscovite contains some iron and magnesia. Winchell's chart suggests that about 60 per cent phengite, 30 per cent Fe‴ muscovite, and 10 per cent pure muscovite are present.

Magnetite is present in appreciable amounts in many of the specimens, from 1 to 2 per cent by volume (Table 1). It is generally in irregular grains, and octahedrons are rare. Magnetite is a characteristic accessory of the Oliverian magma series elsewhere in New Hampshire.

Pyroxene has not been reported previously from the Oliverian magma series. Very small quantities, however, have been found in the coarse granite and coarse syenite. Although sufficient optical data could be obtained to identify the mineral as pyroxene, it was impossible to obtain indices and other significant data.

CONSANGUINITY OF THE GROUP

Two main lines of evidence indicate that the rocks assigned to the Oliverian magma series are truly consanguineous. First, although it has proved entirely feasible to map separate petrographic units, within each there are varieties that simulate other units. Second, the mineralogy of the six rock units are similar in most respects.

The hornblende-quartz monzonite, however, is rather unlike the other rocks of the Oliverian magma series in the Mt. Washington quadrangle. Nevertheless, the differences are largely due to dissimilarities in the percentages of the minerals (Table 1), rather than to any inherent differences in the minerals themselves.

AGE RELATIONS

In the Mt. Washington quadrangle five of the six members of the Oliverian magma series intrude the Ammonoosuc volcanics, but evidence for the sixth member is lacking.

1. Biotite-quartz monzonite. Sills and dikes of the biotite-quartz monzonite cut amphibolite and fine-grained biotite gneiss belonging to the Ammonoosuc volcanics exposed 2 miles south of Black Crescent Mountain (Pl. 1). Similar relations are well

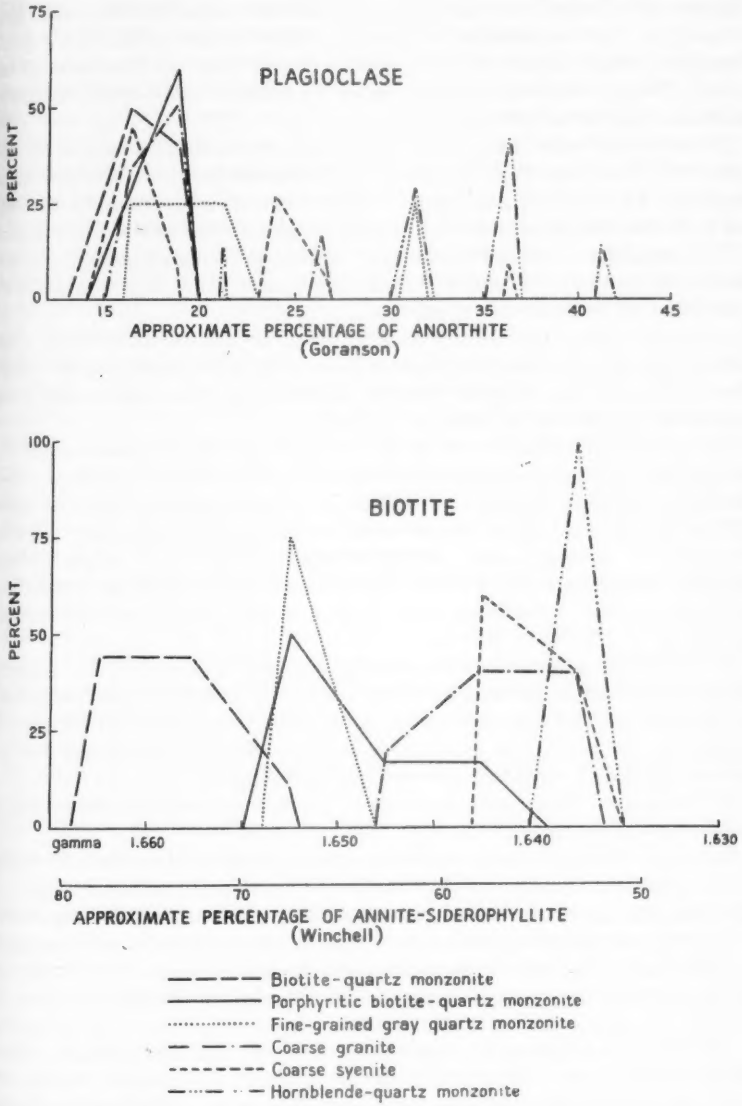

FIGURE 2.—*Frequency diagrams*

Ordinate gives the percentage of all the determinations that fall within abritrary divisions on the abscissa. For plagioclase the abscissa is the percentage of anorthite; for biotite it is the γ index, below which is an additional scale to indicate the approximate percentage of annite-siderophyllite.

displayed on Cold Brook (not labeled on Pl. 1) 2½ miles west-southwest of the village of Randolph. Between altitudes of 1400 and 1480 feet on this brook, biotite-quartz monzonite contains inclusions of amphibolite derived from the Ammonoosuc volcanics. The main body of this formation, cut by a dike of biotite-quartz monzonite begins at an altitude of 1490 feet.

2. Porphyritic biotite-quartz monzonite. Two and a half miles west of Mt. Deception (Pl. 1) a narrow body of Ammonoosuc volcanics is cut by dikes and sills of porphyritic biotite-quartz monzonite. Moreover, inclusions of volcanics are common in the biotite-quartz monzonite along the margins of the Ammonoosuc volcanics.

3. Coarse granite. About 2 miles west of Bowman on the main highway, the coarse granite contains inclusions of fine-grained biotite gneiss which resembles one of the rocks typical of the Ammonoosuc volcanics.

4. Coarse syenite. One mile north of the village of Jefferson an inclusion of amphibolite and several of fine-grained biotite gneiss occur in the coarse syenite. Half a mile northwest of the village of Jefferson, as well as elsewhere, inclusions of fine-grained biotite gneiss are abundant in the syenite.

The intrusive relations are well shown at the contact of the coarse syenite and Ammonoosuc volcanics about 2½ miles northwest of the village of Jefferson. Near the foliation symbol indicating a dip of 60° N., injections of coarse syenite a few inches wide and several feet long are parallel to the foliation of the Ammonoosuc volcanics. Moreover, the volcanics contain numerous large crystals of pink potash feldspar, similar to those in the coarse syenite. In places these large crystals are concentrated in streaks parallel to the foliation and appear to have been introduced from the syenite.

5. Hornblende-quartz monzonite. One-quarter mile southwest of the village of Jefferson Highland, at an elevation of 1330 feet in Stag Hollow Brook (not labeled on Pl. 1), a small slab of biotite-bearing amphibolite, 4 feet wide, is intruded and reworked by the porphyritic hornblende-quartz monzonite. The amphibolite presumably has been derived from the Ammonoosuc volcanics.

6. Fine-grained gray quartz monzonite. No data are available concerning the relative ages of this rock and the Ammonoosuc volcanics.

In the Mt. Washington area the Oliverian magma series could not be dated relative to the formations younger than the Ammonoosuc volcanics (Upper Ordovician?). In the Mascoma quadrangle C. A. Chapman (1939, p. 168) showed that the series is younger than the Silurian Clough formation. On structural evidence Billings (1937, p. 502), C. A. Chapman (1939, p. 146), and Hadley (1942, p. 143) consider the Oliverian magma series to be younger than the Littleton formation (Lower Devonian).

The Oliverian magma series is older than the White Mountain magma series of Mississippian (?) age. Dikes and apophyses of the White Mountain magma series are common in the Oliverian magma series near the contacts. Relations are perhaps most strikingly shown along the "shatter zone," near the villages of Jefferson and Starr King, where coarse syenite has been intensely shattered, sheared, folded, and intruded by many small dikes of the White Mountain magma series. Inclusions of the Oliverian magma series are numerous in the White Mountain magma series.

The ages of the different units of the Oliverian magma series relative to each other in the Mt. Washington quadrangle are not well established.

The hornblende-quartz monzonite appears to be the oldest of the six units in the Oliverian magma series. On the north side of the main highway in the village of Starr King (at the cemetery) and a third of a mile to the northeast (at an altitude of 1600 feet, in a brook) the coarse syenite cuts the hornblende-quartz monzonite.

Just southwest of Jefferson Highland (altitude of 1225 feet on Stag Hollow Brook) a dike of hornblende granite, about a foot wide, striking N. 35° E. and dipping 65° N.E., cuts the hornblende-quartz monzonite. In hand specimen the rock has a fine-grained granular texture, but in the field there is a crude foliation, striking N. 50° E. and dipping 40° NW., essentially parallel to the foliation of the enclosing hornblende-quartz monzonite. This indicates that the stresses that deformed the hornblende-quartz monzonite were operative, at least in part, after the dike had been injected. Microscopic study shows this light-colored rock is composed of microcline, quartz, hornblende, chlorite, and accessories; the mode is given in Table 1. Most of the chlorite has apparently formed from biotite; thus, the fresh material was presumably similar to the coarse granite. The grain size is somewhat finer, however, but the texture is granoblastic like that of the coarse granite. These similarities, together with the fact that the body of coarse granite is less than half a mile distant, strongly indicate that the dike is an apophysis of the coarse granite and that the coarse granite is younger than the hornblende-quartz monzonite.

The hornblende-quartz monzonite is thus older than the coarse syenite and the coarse granite.

The ages of the other units of the Oliverian magma series relative to one another cannot be determined. The biotite-quartz monzonite and the porphyritic biotite-quartz monzonite are so similar that they are probably essentially contemporaneous. The coarse syenite and coarse granite are similar, except for quartz content, and are probably almost contemporaneous.

The chronological arrangement given in the legend on Plate 1 is arbitrary and has no significance.

The Oliverian magma series is considered to be late Devonian (C. A. Chapman, 1942, p. 897).

ORIGIN

The origin of the Oliverian magma series may now be considered. Have all the six mapped units consolidated from magma, or are some of them the product of metasomatic replacement of older metamorphic rocks? Limited replacement associated with the Oliverian magma series has been noted elsewhere in New Hampshire (Billings, 1937, p. 502; C. A. Chapman, 166–176; Hadley, 1942, p. 141).

Most of the mapped units have definitely consolidated from magma. The individual units are relatively large and homogeneous. There are few transition zones from the plutonic rocks into metamorphic rocks, such as we might expect on a replacement theory. Moreover, several of the mapped units occur as sills and dikes in the overlying Ammonoosuc volcanics, or contain sharply delineated, angular inclusions of Ammonoosuc volcanics. The biotite-quartz monzonite, porphyritic biotite-quartz monzonite, and coarse syenite show such relations.

The hornblende-quartz monzonite, an exception to these generalizations, is characterized by heterogeneity. The restriction of the orthoclase "phenocrysts" to definite bands in some localities suggests metasomatic replacement. It is possible, therefore, that some parts of the hornblende-quartz monzonite are replacements of the Ammonoosuc volcanics. But some parts of the hornblende-quartz monzonite seem to have consolidated from magma. One-quarter mile southwest of the village of Jefferson Highland, at an altitude of 1330 feet in Stag Hollow Brook, a porphyritic facies of the hornblende-quartz monzonite contains a small slab of biotite-bearing amphibolite derived from the Ammonoosuc volcanics. This indicates a magmatic origin for the hornblende-quartz monzonite at this locality.

STRUCTURE

GENERAL RELATIONS

The Oliverian magma series of the Mt. Washington quadrangle constitutes the core of a large anticline known as the Jefferson dome, which, elongated in a northeasterly direction, extends far beyond the limits of the Mt. Washington quadrangle (Fig. 1). It is about 42 miles long and about 9 miles wide. As yet, however, detailed mapping has not been done in the Milan, Gorham, and Whitefield quadrangles, and consequently the complete structural picture is unavailable. It may also be seen from Figure 1 that the Jefferson dome is in line with several other domes and that it is by far the largest in the group. In all these domes the Oliverian magma series has intruded metamorphosed Ordovician (?) rocks and has displaced the overlying formations.

STRUCTURE OF SURROUNDING METAMORPHIC ROCKS

In the Mt. Washington quadrangle a discontinuous belt of Ammonoosuc volcanics occupies the southeast flank of the plutonic core. Southeast of and overlying these volcanics is the Partridge formation. The bedding and foliation of both formations, although somewhat folded, dip southeast (Pl. 1). The southeast flank of the dome is broken by the Pine Mountain fault, already described by Billings (1941, p. 914). The fault extends for 13 miles in a northeasterly direction and follows a line of silicified zones. The dip of the fault is steep, and the southeast side is downthrown. Intrusives of both the New Hampshire and White Mountain magma series further disrupt the continuity of the southeast flank of the dome.

Reconnaissance shows that the belt of Ammonoosuc volcanics bounding the plutonic mass on its southeast flank extends northeast for several miles into the Gorham quadrangle (Fig. 1). In the southwest part of the Mt. Washington quadrangle and in the adjacent Whitefield and Franconia quadrangles, the Ammonoosuc volcanics have been cut out by intrusives belonging to various magma series. In the western part of the Franconia quadrangle, two patches of vertical Ammonoosuc volcanics (William and Billings, 1938), south of the Scrag granite (now correlated with the Oliverian magma series), are remnants of the belt of Ammonoosuc volcanics bounding the pluton on the southeast.

The Ammonoosuc volcanics 4½ miles north of the village of Jefferson lie on the northwest flank of the dome, dip 50° NW., and extend for several miles southwesterly into the Whitefield quadrangle. For 8 miles to the northeast the Ammonoosuc vol-

canics on the northwest flank of the dome are cut out by intrusives of the White Mountain magma series, but they reappear in the Percy quadrangle, dipping 50° NW. (R. W. Chapman, 1935, p. 405) and continue for several miles into the Milan quadrangle (Fig. 1).

In the northwest corner of the Mt. Washington quadrangle the Ammonoosuc volcanics are separated from the Albee formation and the Lost Nation group by what appears to be a fault. Reconnaissance in the Whitefield quadrangle (Fig. 1) indicates that this fault is a continuation of the Ammonoosuc thrust of the Littleton quadrangle (Billings, 1937, p. 525).

In summary, it appears that Ammonoosuc volcanics overlie both flanks of the intrusive mass and at one time may have formed a complete ring about the plutonic core. In this respect, therefore, the Jefferson dome resembles the other domes to the southwest (Billings, 1937; C. A. Chapman, 1939; 1942; Hadley, 1942). Faulting and later igneous intrusions, however, have obliterated large portions of this ring of metamorphic rocks.

STRUCTURE OF IGNEOUS CORE

The attitude of the foliation of the igneous core of the Jefferson dome indicates a huge anticline trending northeast. The dome is asymmetric; the northwest limb has an average dip of about 45° NW., whereas the southeast limb has an average dip of about 75° SE. and locally shows overturned, steep, northwesterly dips. Moreover, the belt of intrusive rocks on the southeast limb is about 3 miles wide, whereas that on the northwest limb is at least twice as broad. Thus the "thickness" of the plutonic rocks on the southeast limb is about 10,000 feet, whereas the "thickness" on the northwest limb is about 20,000 feet. (*See* the structure section, Pl. 1.) In detail, the igneous core of the dome is composed of several lithologic units. Apparently each unit is more or less lens-shaped. This is suggested by three facts: (1) The units are elongate in ground plan; (2) the primary foliation of each unit parallels the contacts of that unit; and (3) the longer dimensions of each of the petrographic units closely parallel each other.

The dome is thus actually a composite body, made up of several overlapping lenses. Some of the petrographic units may themselves be composite, composed of separate injections.

NATURE OF INTRUSION

The entire Jefferson dome should be studied in detail before its origin can be completely understood. The writers believe, nevertheless, that the general sequence of events is clear.

The Oliverian magma series is believed to have intruded the Ammonoosuc volcanics, more or less conformably. In these respects the Jefferson dome resembles the other Oliverian domes. The composite igneous core is believed to have formed by the forceful intrusion of several successive lenticular magmatic bodies.

That the magmas were injected forcefully is suggested by flow structures that are present throughout the igneous rocks. Foliation due to oriented biotite flakes and hornblende crystals is nearly everywhere present. Large feldspar crystals in parallel arrangement give much of the coarse syenite a trachytoid texture. Parallelism of

large feldspar crystals is observed locally in the coarse granite and porphyritic biotite gneiss. Slablike inclusions, not necessarily confined to contact zones, but widely scattered through the igneous rocks, are invariably parallel to the foliation. No screens of older metamorphic rocks have been recognized between the different intrusions, unless, as discussed on a preceding page, some parts of the hornblende-quartz monzonite are altered Ammonoosuc volcanics.

Similar domes occur in northern Massachusetts, due south of the Bellows Falls quadrangle (Fig. 1). The Pelham dome is composed of concordant sheets and sills injected into sedimentary gneiss (Balk, 1942). The pre-Cambrian dome around Harney Peak, Black Hills, South Dakota, is also a structural complex composed of many sills (Runner, 1943).

How much of the anticlinal structure of the Jefferson dome is primary, inherited from the time of intrusion, and how much is secondary, due to post-consolidation compressive forces, cannot be stated definitely.

CORRELATION WITH OLIVERIAN MAGMA SERIES ELSEWHERE

These igneous rocks of the Mt. Washington quadrangle may be correlated with the Oliverian magma series in quadrangles to the southwest (Fig. 1) for the following reasons: (1) mineralogical and textural similarity; (2) similarity in age; and (3) similarity in structure.

The biotite-quartz monzonite is similar to the Owls Head granite of the Moosilauke quadrangle (Billings, 1937, p. 501) and Rumney quadrangle (Fowler-Billings and Page, 1942), the Smarts Mountain granodiorite of the Mt. Cube quadrangle (Hadley, 1942, p. 137), and the quartz monzonite and granodiorite of the Mascoma and Croydon groups in the Mt. Cube, Mascoma, and Sunapee quadrangles (Chapman, 1939, p. 143). The porphyritic biotite-quartz monzonite is similar to the granite and quartz monzonite of the Mascoma group in the Mascoma quadrangle (Chapman, 1939, p. 143). The fine-grained, gray quartz monzonite resembles some of the finer-grained rocks of the Mascoma and Croydon groups (Chapman, 1939, p. 143). Much of the coarse granite is similar to the pink granite of the Mascoma group (Chapman, 1939, p. 143). The essential mineralogical difference is the presence of hornblende in the coarse granite.

In most respects the coarse syenite and hornblende-quartz monzonite are unlike any previously described rocks in the Oliverian magma series. The coarse syenite, moreover, is superficially similar in its mineralogical composition to the syenites of the White Mountain magma series but differs in several ways: (1) The pink color of the potash feldspar is more intense than in the White Mountain magma series; (2) the characteristic planar parallelism of the well-formed tabular crystals of potash feldspar is unknown in the White Mountain magma series; (3) the microperthite of the coarse syenite shows relatively little albite included in the potash feldspar compared with the microperthite of the White Mountain magma series; and (4) inclusions in the coarse syenite are nearly everywhere oriented, which is rarely true in the White Mountain magma series. Perhaps the most important reason for not assigning the coarse syenite to the White Mountain magma series is the fact that the coarse syenite is older than all members of this series in the area. The numerous studies of the

White Mountain magma series have well established the rule that the syenites are younger than the more mafic and less alkalic types (Chapman and Williams, 1935). In the present area, however, inclusions of the coarse syenite of the Oliverian magma series are found in the most mafic phases (quartz monzodiorite) of the White Mountain magma series, and dikes of the latter cut the coarse syenite, particularly along the "shatter zone." It seems very certain, therefore, that the coarse syenite cannot be grouped with the White Mountain magma series.

The structural similarity of the Jefferson dome to other domes in western New Hampshire is apparent. All the domes have a core of plutonic rocks and a cover of Ammonoosuc volcanics and younger rocks. Some of the domes are symmetrical, with opposite limbs dipping off the dome at similar angles; the Owls Head, Croydon, and the Unity domes are of this type. Other domes are asymmetrical and overturned toward the west; the Smarts Mtn. and Mascoma domes are of this type, The Jefferson dome is unusual in that the axial plane dips north, because the southeast limb is steeper than the northwest limb.

The arguments that the plutonic cores of the domes were emplaced prior to the main period of folding has been presented elsewhere (Billings, 1937, p. 535–536; C. A. Chapman, 1942, p. 911, 915).

SUMMARY

Rocks belonging to six mappable petrographic units constitute the late Devonian (?) Oliverian magma series in the Mt. Washington quadrangle. These rocks are pink to gray, porphyritic to equigranular, and range from coarse-grained to fine-grained. Foliation, produced by parallelism of biotite, hornblende, potash feldspar, or quartz, is common. The petrographic types represented are: (1) hornblende-quartz monzonite, (2) fine-grained gray quartz monzonite, (3) coarse syenite, (4) coarse granite, (5) biotite-quartz monzonite, (6) porphyritic biotite-quartz monzonite.

Direct field evidence shows that the coarse syenite and coarse granite are younger than the hornblende-quartz monzonite. Direct evidence of the relative ages of the other petrographic units, however, is lacking.

The petrographic types are correlated with rocks of the Oliverian magma series, elsehwere in New Hampshire, for several reasons: (1) similarity in mineralogy and textures; (2) similarity in time of intrusion; and (3) similarity in structure.

Many tens of square miles of the Oliverian magma series have been mapped in New Hampshire, and in all cases the rocks lie within oval-shaped areas, distributed along a belt trending northeast for over 125 miles. In each case the plutonic mass within the oval-shaped area is a structural dome surrounded by stratified Ordovician (?) and younger rocks. In the Mt. Washington quadrangle the Oliverian magma series forms part of the Jefferson dome, which, inasmuch as it is 42 miles long and 9 miles wide, is the largest dome of the group. Perhaps less than half the dome has been studied in detail, but reconnaissance in adjacent quadrangles has established the general structure.

The best evidence available indicates that the dome was formed by a series of lenticular intrusions that formed a composite igneous body.

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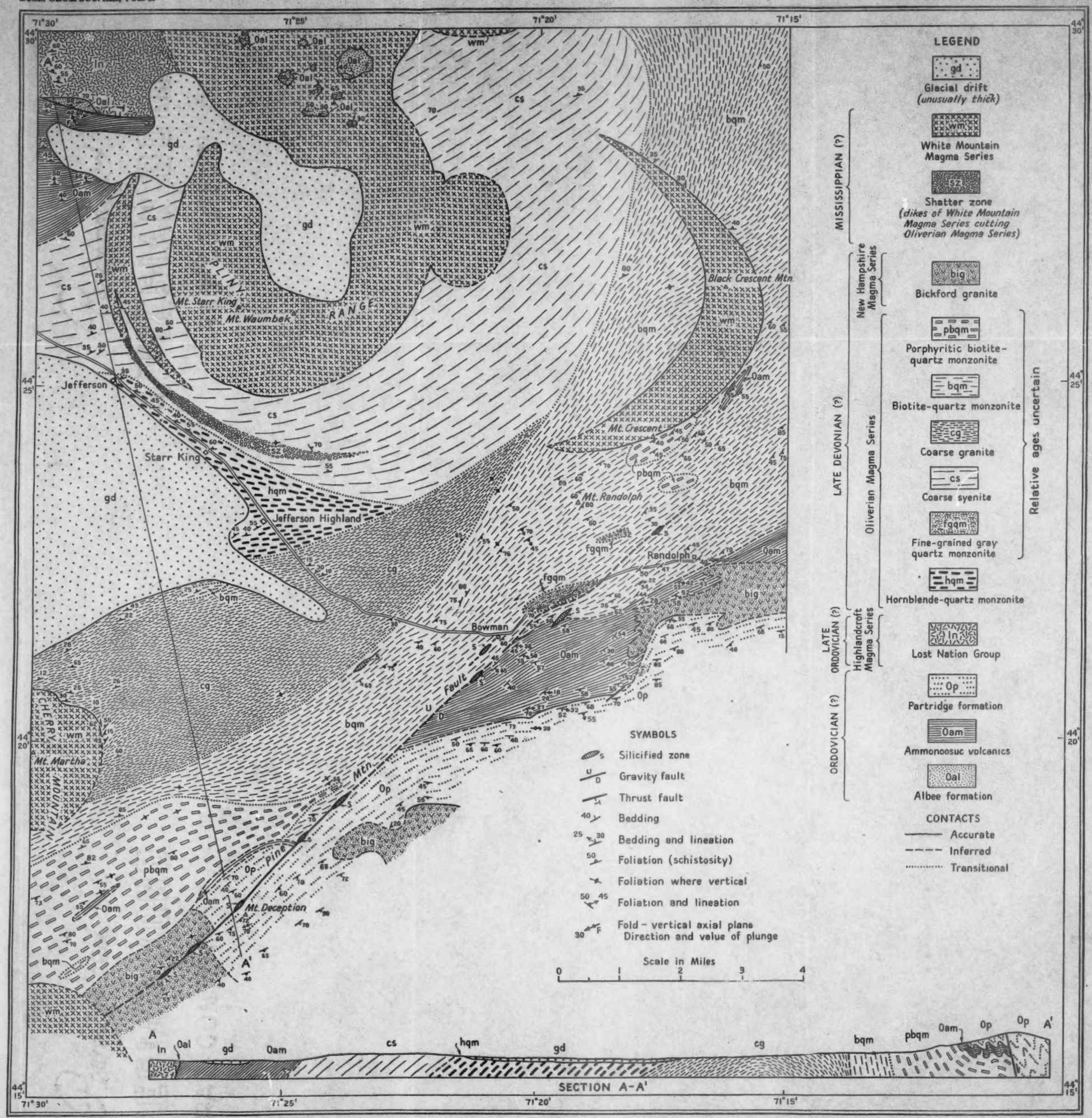

GEOLOGIC MAP AND CROSS SECTION OF PART OF THE MOUNT WASHINGTON QUADRANGLE, NEW HAMPSHIRE